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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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GLACIAL GEOLOGY OF GRAND COULEE DAM, WASHINGTON

BY RICHARD FOSTER FLINT AND WILLIAM H. IRWIN

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ABSTRACT

Sections of the fill in Columbia canyon near Grand Coulee Dam, together with the stratigraphy and morphology of related features downstream, record the expansion and later shrinkage of the Okanogan glacier lobe during the last glaciation. Three stratigraphic zones are recognized: (1) A Basal Sequence of lacustrine fines interbedded with fluvial sediments, recording oscillatory ponding of the Columbia River as the glacier dammed it downstream. At first the lake was shallow and drained west around the ice margin. Later it was more effectively dammed, deepened, and forced to discharge south through Grand Coulee. (2) A Till Zone, recording the arrival of ice, with fluctuations, at Grand Coulee Dam during the glacial climax. (3) Later deposits stratigraphically (but not everywhere vertically) overlying the Till Zone, recording oscillatory deglaciation accompanied by transition from lacustrine to fluvial conditions.

The San Poil glacier lobe entered the lake-filled Columbia canyon at Keller Ferry, 15 miles upstream from Grand Coulee Dam, and contributed sediment to the lake. The maximum of this lobe appears to have been approximately contemporaneous with the maximum of the neighboring Okanogan lobe.

While residual ice was still present locally in Columbia canyon downstream, drainage abandoned Grand Coulee and resumed its former (and present) course. Thereafter, grading of the long profile of Columbia River proceeded both by erosion of the lake fill and by deposition in depressions partly of ice-block origin and was marked by at least one delay recorded by a particularly conspicuous stream terrace.

INTRODUCTION

The unusually complete and detailed section of the glacial fill fully exposed by excavations during the construction of Grand Coulee Dam and the glacial features in the region of the dam combine to furnish a far more complete picture of the continuous expansion and shrinkage of a large glacier than could have been reconstructed from natural exposures alone. It was thought desirable to interpret and correlate the features and to present the results in condensed form, not only as a contribution to the glacial history of eastern Washington, but as a yardstick that might prove useful for comparison with somewhat similar but less well-exposed glacial sections elsewhere.

Irwin, as assistant to Dr. Charles P. Berkey and Junior Geologist in the United States Bureau of Reclamation, spent 4 seasons (summers 1935-1938) studying the general geology of the Grand Coulee Dam region. Part of this time was given to the study of the glacial deposits at and near the dam.

Flint, with a background of 4 seasons' study of the glacial deposits of this general region and visits to the dam during the earlier stages of construction, spent 2 weeks in July 1938 in the vicinity of the dam and 2 weeks farther downstream. Flint and Irwin visited together the critical localities at and near the dam. They are jointly responsible for interpretation, worked out in preliminary form by Irwin, of the features at these places; Flint is solely responsible for inferences drawn from localities more remote from the dam.

ACKNOWLEDGMENTS

The authors wish to express their appreciation to the United States Bureau of Reclamation, through Mr. R. F. Walter, Chief Engineer, for permission to publish Reclamation data and photographs; and to Mr. Frank A. Banks, Construction Engineer for the United States Bureau of Reclamation at Grand Coulee Dam, for the opportunity afforded the junior author of conducting the investigation. It is a pleasure to acknowledge the active interest and cooperation of Mr. Banks and the members of his staff. Sincere thanks are due to Professor Charles P. Berkey of Columbia University for making available memoranda prepared by him as Consulting Geologist for the United States Bureau of Reclamation in addition to his personal observations made during the initial stages of the excavation operations. Acknowledgment is hereby given of a grant from the Penrose Bequest of the Geological Society of America which made possible the work of the senior author during the season of 1938. Dr. Paul D. Krynine of Pennsylvania State College kindly made a mechanical and mineralogic analysis of a sample of silt from the Basal Sequence, with the results briefly outlined in the text. The efficient cooperation of the 116th Photo Section, Washington National Guard, through Mr. Miller Cowling, Chief Scientific Aid, in making aerial photographs has been especially helpful. Finally, the senior author cordially acknowledges the work of his field assistants, Mr. M. B. Canning and Mr. W. E. Benson, both of Yale University.

ADVANCE SUMMARY

The Columbia River (Fig. 1) follows a broadly sinuous course along the northern margin of the Columbia Plateau and occupies a capacious canyon approximating 1500 feet in depth. The elevation of the river surface at mean low water is 980 feet at the mouth of the San Poil River, 940 feet at Grand Coulee Dam, 900 feet at the mouth of the Nespelem River, 860 feet opposite the Omak Lake Trench, and 740 feet at the mouth of the Okanogan River. The canyon contains extensive remnants of a partial sedimentary fill made during the last Pleistocene glacial stage. It has been known for many years that the Columbia River was dammed by glacier ice and that the Grand Coulee served as a temporary spillway for the impounded water. A detailed study of the fill has permitted reconstruction of the details of these events. The section exposed at Grand Coulee Dam (Pl. 2; *see also* Irwin, 1938) records the blocking, at first temporary and unsuccessful, of the Columbia River, followed by effective damming, extensive sedimentation in the lake thereby formed, and advance of the glacier margin to an extreme position east of the head of Grand Coulee, diverting the outflow through the coulee. The record

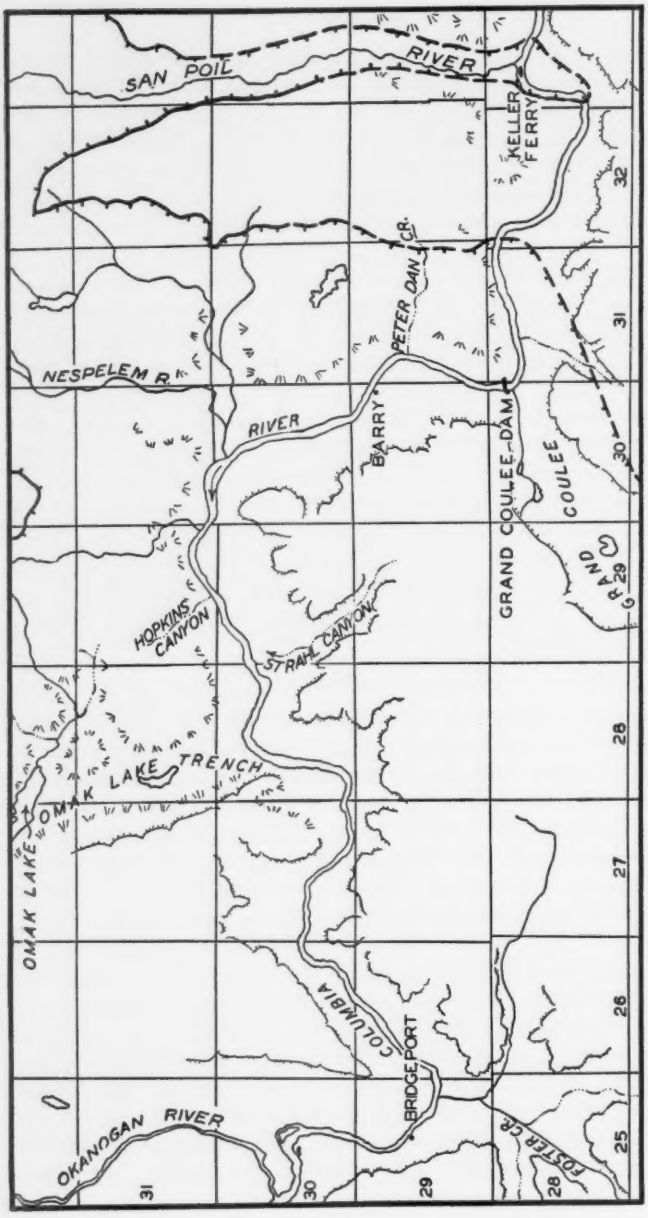


FIGURE 1.—*Map showing regional relations of Grand Coulee Dam*

Scale: one square equals approximately 6 miles on a side. Thin hatched line — approximate position of prominent basalt cliffs. Hachures — granitic rock slopes along Columbia River. Thick hatched line — inferred drift border (Flint, 1937, Pl. 5, with later addition near Keller Ferry). Locations of places and features mentioned in the text are shown, and townships are numbered for reference.

further shows that deglaciation was accompanied by sedimentation in the wake of the shrinking glacier, abandonment of the temporary Grand Coulee discharge route, and resumption of the drainage course now followed by the Columbia River.

DESCRIPTION OF THE FILL SECTION

SOURCES OF DATA

Plate 1 is a simplified cross section illustrating the significant features of lithology and structure typical of the valley fill at the site of Grand Coulee Dam. The line of the section, about 5000 feet long, passes from the east abutment across the floor of the dam to the vicinity of the downstream side of the west abutment and thence northwestward along the axis of a tributary ravine [1] just north of the west abutment. Although the deposits illustrated represent a total vertical thickness of about 800 feet, the actual thickness of material excavated at any one point varied roughly from 50 to 300 feet.

The data have been assembled from a study of the records of many exploratory borings and from the personal observations of the junior author and Dr. C. P. Berkey and others over a period of about 4 years. In the progress of excavation operations during this period, various parts of the section were successively exposed and destroyed, and at no one time were very large portions continuously available for critical study. The interpretations presented and implied in the cross section have been checked by field observations and by the preparation and study of 12 additional cross sections along lines parallel with or transverse to the approximately east-west axis of the dam.

WEST SIDE OF RIVER

General statement.—For clarity in discussion the fill section west of the river will be divided into three major groups: (1) A basal sequence of predominantly glacial varved silt and clay, overlain by (2) an intermediate zone of till and associated poorly stratified silt, sand, and gravel, overlain by (3) stratified glacial-lacustrine silt and glaciofluvial sand and gravel.

Basal Sequence.—The material of the Basal Sequence (Pl. 3, fig. 1) is composed predominantly of a light-gray lacustrine silt, tough and hard when dry but decidedly plastic when wet. Clay is mixed with the silt in variable amounts. The silt is in horizontal beds characterized either by varves or by an almost total lack of discernible stratification. The varved silt layers vary from about one foot to 14 feet in thickness, and the massive silt layers from a few inches to 9 feet. There are a few beds of fine to

[1] Marked in Plate 2, figure 1, by a large semicircular artificial excavation.

coarse sand, some of which carry small amounts of pebble gravel. They vary from a fraction of an inch to 3 feet in thickness.

As illustrated in the cross section (Pl. 1), the lowermost bed in the Basal Sequence consists of a massive light-brown silt layer about 6 feet thick showing only slight evidence of current action. Wherever observed it rested directly on the water-worn granite floor of the preglacial Columbia River with the exception that, in open crevices and narrow linear depressions extending beneath the general bedrock surface,[2] the silt was seen to be underlain by meager amounts of sand and gravel.

Above this massive silt bed is a series consisting of massive to poorly varved silt and thin beds of sand and granule gravel. These succeeding strata are marked chiefly by their irregularity in relative thickness, in succession of types, and in perfection of internal structures. This dominant irregularity fails with the gradual appearance of a repeated sedimentary cycle: (1) Sand, (2) massive silt, (3) varved silt. The sand member of this three-part sequence grades up from coarse, current-bedded sand, in places gravel-bearing, into finer sand.

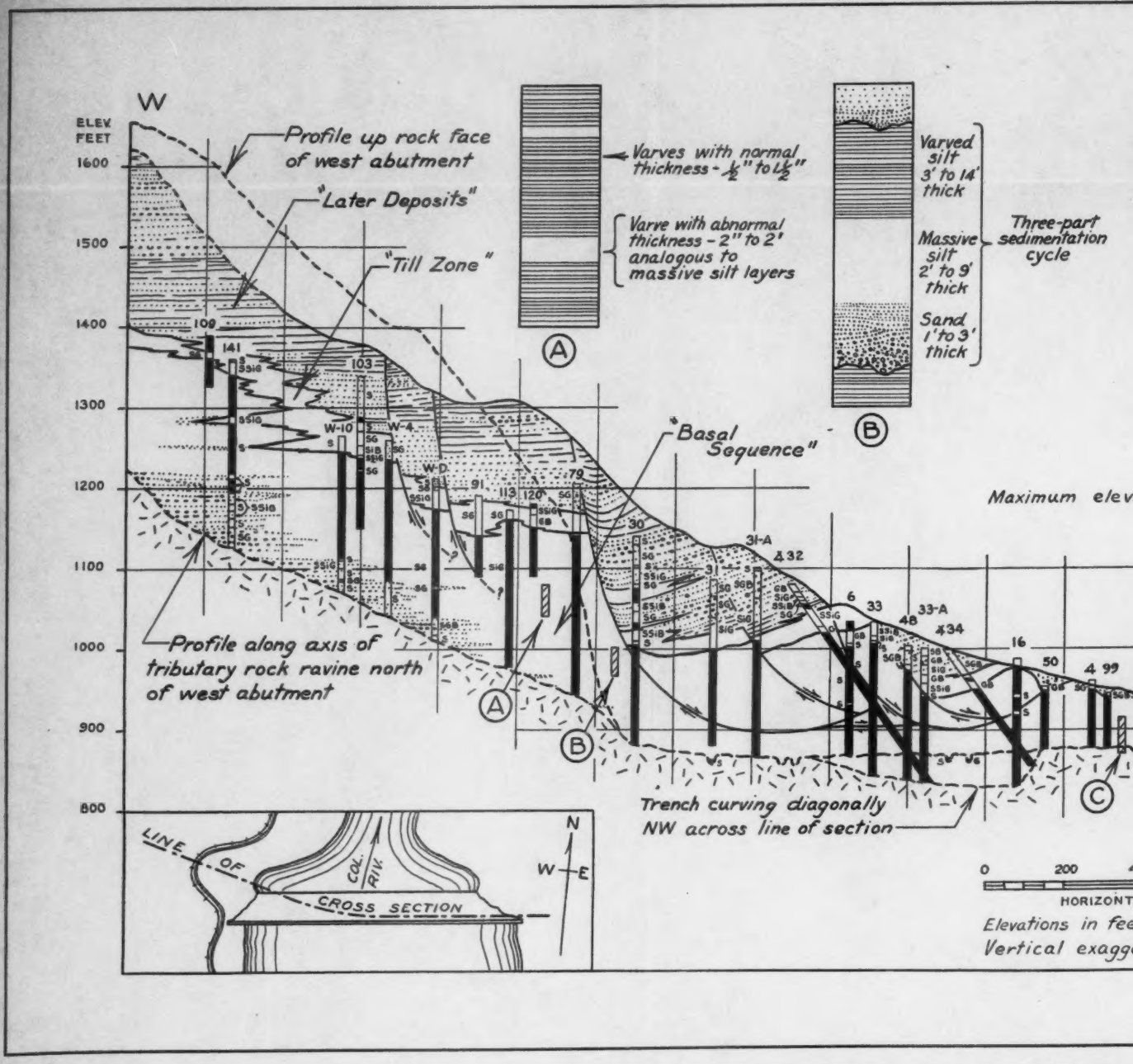
The middle member, composed of silt, is coarse grained and usually faintly current bedded near its base, and grades up into fine-grained nonstratified silt. The upper member consists of definitely varved silt. The contacts between the three members of any sequence are gradational, but the sand member commonly rests with slight disconformity upon the varved member of the preceding sequence. In none of the successive disconformities between cycles was any evidence seen that the top of the varved silt member had been eroded more than a few inches.

Higher in the section, the sand member decreases in thickness and finally disappears entirely. The massive silt member likewise decreases in thickness and grain size, with the result that the varved structure predominates. At irregular intervals there occur abnormally thick varves. Still higher, the distinctly varved silt grades into thin-bedded laminated coarser silt and very fine sand. The occurrence of interbedded coarse stratified sand and gravel with increasing frequency marks the change from the Basal Sequence into the overlying till zone.

The following data have been abstracted from a report kindly made by Dr. Paul D. Krynine on a silt sample from the Basal Sequence west of the river:

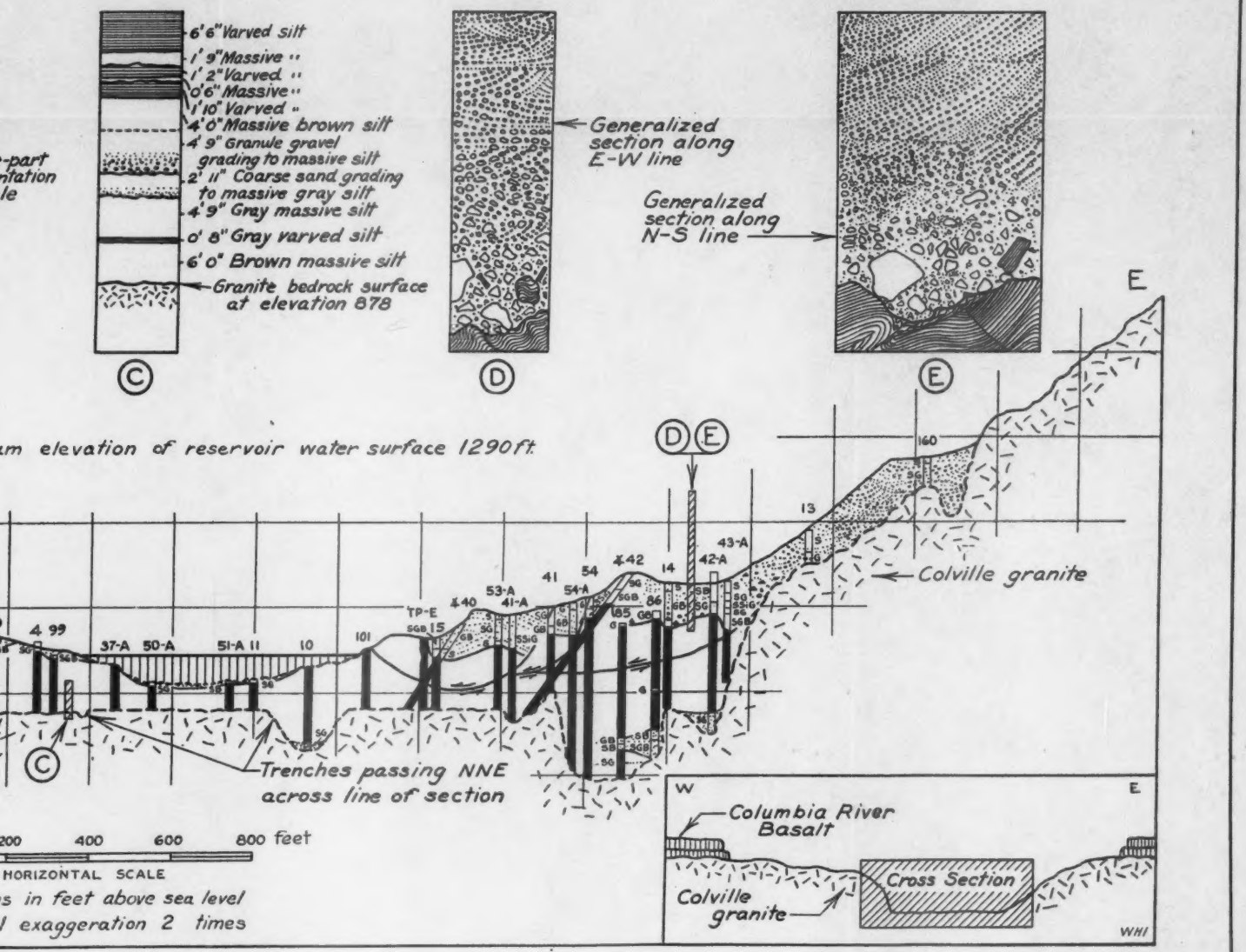
> The sample is an olive-greenish-gray, finely laminated clayey silt. Alternate dark and light layers suggest varving and vary between 0.1 mm. or less, and 0.7 mm. in thickness. Only 1.24 per cent of the material is coarser than 0.044 mm. It exhibits notably heterogeneous size sorting, degree of attrition, and alteration.
>
> The coarse fraction is composed of quartz, potassium feldspar and sodic to calcic plagioclase, fragments of gneiss, basalt and greenstone, mica, andesite, and volcanic

[2] The attention of the reader is directed to the unusually even profile of the rock floor in the western and central parts of the preglacial channel (Pl. 1).



CROSS SECTION OF THE FILL IN COLUMBIA

Narrow vertical columns show actual or projected positio material encountered shown by symbols: S—sand; Si—silt; black. Individual drawings (labeled A to E) above cross profi cross profile represented by vertical, diagonally cross-hatched slumps indicated by curved lines. Distribution of silt, sand and and boundaries of Till Zone based on interpretation and field ob



LUMBIA RIVER CANYON AT GRAND COULEE DAM

ted positions of drill holes along line of cross profile. Driller's log of Si—silt; G—gravel; B—boulders; massive or varved silt shown in cross profile illustrate typical details at similarly labeled positions in -hatched columns. Approximate position of shear surfaces of major , sand and gravel (represented by lines and dots of appropriate sizes), nd field observations (Constructed by William H. Irwin).

glass (listed in approximate order of abundance). Heavy minerals, which constitute 8.6 per cent of the coarse fraction: hornblende, olivine, garnet, kyanite, hypersthene, zircon, epidote, tourmaline (listed in order of abundance); traces of barite, sillimanite, staurolite, zoisite(?); opaque "iron ores" (*i.e.*, magnetite, ilmenite, and leucoxene) and undeterminable dark ferruginous aggregates.

The fine fraction is composed of about one-third finely comminuted quartz and the minerals mentioned in the coarse fraction, and about two-thirds clay substance. The latter is a brownish-gray, micaceous, weakly anisotropic paste almost irresolvable under the microscope.

Till Zone.—The central portion of the Till Zone consists of a continuous layer of till (Pl. 3, fig. 2) which varies in thickness. The poorly sorted sand and gravel both above and below the till contain lenses of till-like debris and some beds of laminated lacustrine silt. As the cross section indicates, portions of the till layer repeatedly finger out into adjacent glaciofluvial gravels, giving an erroneous impression in restricted exposures of two distinct layers of till.

Through several miles downstream from the dam, natural exposures likewise reveal that a zone of till similar to that described above separates the varved silt of the Basal Sequence from the overlying stratified lacustrine silt. The varves of the Basal Sequence are in many places markedly distorted and crushed.

Later deposits.—West of the river the Till Zone grades up into relatively little compacted stratified lacustrine silt. In places this silt is interbedded with layers of sand and gravel derived from side-valley drainage. This silt is correlative with the Nespelem silt (Pardee, 1918, p. 28, 47), although lithologically it is not wholly typical at the immediate site of the dam. The Nespelem silt is a late-glacial, lacustrine valley fill whose remnants extend as terraces through many miles upstream and downstream from the dam (Pl. 4). The terraces have been described in previous publications (Flint, 1935; 1936) and are mentioned elsewhere in the present paper.

EAST SIDE OF RIVER

As indicated in the cross section (Pl. 1), the varved silt discussed under "Basal Sequence" in the fill section west of the river is continuous across the valley floor and underlies all the deposits east of the river. Here the remainder of the fill section is composed of till and deltaic deposits both of which are referred to hereafter as simply "deltaic deposits." The top of the fill section is the surface of a prominent stream terrace which extends through considerable distances along the east side of the valley. The deltaic deposits have been thus far identified solely in the fill east of the river, where they constitute the upper portions of the prominent stream terrace [3] from a point about three-fourths of a mile above the dam to a point about 1½ miles downstream. The maximum thick-

[3] Mason City, the contractor's town, is situated on this terrace.

ness of the deltaic deposits is about 200 feet with their present upper limit between elevations of 1200 and 1225 feet. The original upper limit is not known but may not have been more than 50 feet higher.

The deltaic deposits rest with marked unconformity upon the irregular surface of greatly distorted varved silt and upon essentially nonstratified disturbed coarse silt. They are overlain by a surficial veneer of roughly horizontally stratified fluvial sand and gravel, from a few feet to 30 feet thick, deposited by the river while engaged in cutting the terrace surface by lateral planation.

Materials of the deltaic sequence show great vertical variation in size, structure, and lithology. The sequence consists predominantly of sand and gravel above, grading irregularly down into till. The gravel is composed of considerably rounded pebbles and cobbles. Delta foresets (Pl. 5, fig. 1) dipping downstream and showing successive cut-and-fill variations are developed in it but become indistinct toward the base. Horizontally stratified sand and pebbles are associated in places with the foreset beds. The till is a rubbly mass of angular stones. It carries a miscellaneous assortment of granites, basalts, metamorphic rocks, fragments of crushed and distorted varved silt, large blocks of massive silt, fragments of the Latah formation, and till balls. The gravel above the till carries basalts, granites, metamorphic rocks, and considerable quantities of andesitic and rhyolitic lavas with several different lithologies. The gravel lacks the more friable (Latah and glacial) constituents. However, the rhyolitic and andesitic fragments are almost completely lacking in the till, become increasingly numerous toward the top of the deltaic deposits, and are quite common in the still younger veneers of gravel on the terrace surfaces. This fact is believed to have an important bearing on the genesis of the deltaic deposits.

LANDSLIDES

Excavations revealed that a major portion of the valley fill, except that close to the bedrock floor or in a protected position near the canyon walls, was somewhat disturbed. The degree of the disturbance varied from simple tilting to complex faulting and mashing of originally horizontal beds. It was due partly to the thrust of advancing ice but predominantly to the slumping of materials which had been either left in unstable positions by removal of ice support during deglaciation or undercut by river erosion. Disturbance of the materials by excavation operations resulted in new movements along both old and new slip-surfaces. The approximate positions of the more prominent slip-surfaces have been indicated in Plate 1. What little evidence of slumping there was in topography and natural exposures gave no indication of the actual severity and frequency of the movements.

FIGURE 1. VIEW SOUTH (UP-RIVER)
Showing stream terraces cut from glacial fill and hanging intake of Grand Coulee. Trestle in background represents position of dam. (Photo by 116th Photo Section, Washington National Guard.)

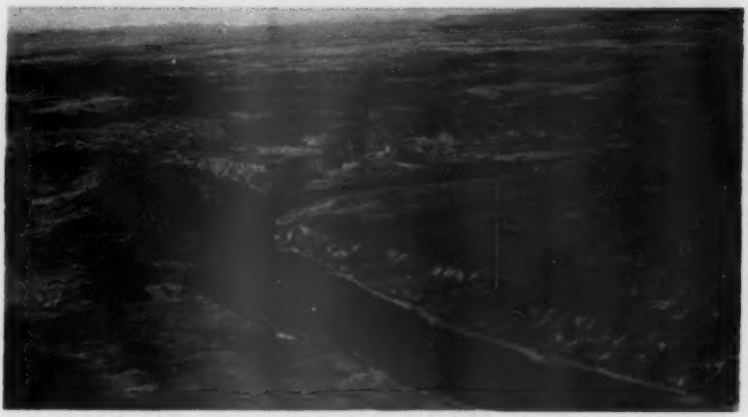


FIGURE 2. VIEW WEST (DOWN-RIVER)
Showing hanging intake of Grand Coulee (left) and a stream terrace cut from the glacial fill (here the Nespelem silt) (right). (Photo by the 116th Photo Section, Washington National Guard.)

GRAND COULEE DAM

FIGURE 1. BASAL SEQUENCE
Mild distortion in beds of varved and massive silt west of river.

FIGURE 2. TILL ZONE
Illustrates till at base of deltaic sequence resting on massive silt in east bank of river about 2000 feet south of dam.

FILL SECTION AT THE DAM

MORPHOLOGY OF THE FILL

NESPELEM SILT TERRACE

The glacial deposits at the dam and along the Columbia River both east and west of the dam occur as terrace remnants. The Nespelem silt terrace may be used as a convenient reference datum because it is conspicuous both up- and downstream from the dam although it is not prominent near the dam itself. This feature (Pl. 4) is a paired terrace cut from lacustrine filling of the canyon. Its surface represents the surface of lacustrine upbuilding (with minor fluvial contributions) at the time when the drainage abandoned the Grand Coulee route and returned to its present route.

HIGHER, ICE-MARGINAL TERRACES

Tributary entrants both large and small off the Columbia canyon contain local terracelike fills lying higher than the Nespelem silt terrace. They are present along both sides of the Columbia River near the head of Grand Coulee at elevations between 1700 and 1900 feet. Farther downstream they occur at elevations up to more than 2500 feet, probably owing to increasing elevation of the upper surface of the ice toward the north. These terracelike fills are believed to be ice marginal because they consist of sand, gravel, and boulders that include both rocks derived from the north and stones with glaciated shapes, and because some of them are kettled. A fill at an elevation of about 1900 feet, in a broad entrant in secs. 13 and 14, T.29N., R.30E., near Barry, is an extensive mass of knolls and kettles.

These terracelike fills date from a time when the glacier was at and near its maximum. They are younger than the Basal Sequence at the dam and older than the silt and other sediments near the top of the fill section at the dam.

The concrete aggregate used in the construction of the dam and appurtenant structures is being obtained from one of these fills, locally named the Brett gravel pit. This fill is located in an entrant in the east wall of the valley about 1½ miles north of the dam. Its terracelike surface approximates an elevation of 1900 feet and exhibits several undrained depressions and many large basalt erratics. Its materials (Pl. 5, fig. 2) include some till, with larger quantities of sand, gravel, and boulders, showing a wide range of size-sorting and lithology and stratification that predominantly dips very gently southeast into the valley wall.

LOWER, STREAM TERRACES

The fill at elevations lower than the surface of the Nespelem silt terrace occurs as discontinuous stream-terrace remnants. These nonpaired benches (Pl. 4) are commonly surfaced with rounded gravel derived in

part from up the Columbia River and are believed to be the result of dissection by the Columbia River shortly after it abandoned the Grand Coulee route and resumed its former course.

INTERPRETATION OF FILL SECTION

GENERAL STATEMENT

Detailed study of the complete valley fill section at the site of Grand Coulee Dam and of associated deposits near the dam yields evidence of only one major glacial advance and retreat, accompanied by minor ice-front oscillations. The various parts of the valley fill form a record, continuous on the west, but discontinuous on the east side of the river, of ponding of the Columbia River as the glacier blocked it downstream; subsequent advance of the ice over, and a short distance upstream from, the site of the dam; and several phases of ensuing deglaciation.

WEST SIDE OF RIVER

The Basal Sequence of the sedimentary fill is a record of ice expansion, which created a lake whose waters ultimately drained through the high-level spillway afforded by the Grand Coulee. The repeated occurrence of the three-part sequence—sand, massive silt, varved silt—reflects cyclic variations in the conditions of sedimentation due to changes in absolute depth of the lake and to fluctuations in the amount of sediment delivered. Marked shoaling of the lake, even to the extent of causing the temporary return of river flow, is indicated by the beds of sand and the slight disconformities that commonly separate them from the finer-grained deposits that immediately underlie them. Although no channeling of the varved silt was seen, the presence of current-bedded sand and granule gravel indicates the existence of a definite current at the time of their deposition.

The ponding at Grand Coulee Dam was caused by the Okanogan glacier lobe crossing the west-flowing segment of the Columbia River at some place between the mouth of Nespelem River and the mouth of Okanogan River. The general southerly movement of the ice in this section was transverse to the trend of the canyon. Hence, as the ice front moved into and across the valley the lake level must have reflected any variation in the rate of ice advance [4] or height of the ice front. In addition, before a complete and final blockade of the canyon was effected, there must have been times when the obstruction was high enough to raise the lake level and result in overflow southwest across the plateau through the high-level Grand Coulee spillway. That the Grand Coulee intake was already in existence at this time is inferred from evidence presented elsewhere in the

[4] Variation in the rate of ice-front advance, resultant variation in the channel cross section through which the normal river flow passed, and the concurrently varied erosive power of the water as it attempted to remove the ice obstruction must all have combined to effect repeated changes in the lake level.

present paper. At such times, conditions of sedimentation must have been sufficiently stabilized to permit the formation of varved silt and possibly also of massive silt. Likewise there must have been times when reduced rate of ice advance and resultant lowering of the ice dam would have permitted temporary resumption of low-level drainage across the ice obstruction along the normal route of the Columbia River. Shallow-water conditions and even the return of seminormal river flow would have resulted in the formation of massive silt and sand beds.

Changes in the quantity of material transported into the lake must have operated also to vary the sedimentary record. It is, however, more difficult to identify the results of such changes than it is to identify those due to lake-level fluctuation. A consideration of the geographic and topographic conditions shows this situation to be quite expectable. The moderate to strong relief of the land surface north of the Columbia River provided ample opportunity for creation of temporary high-level lakes dammed by the ice front or by glacial deposits. The filling up of such a lake or its sudden drainage due to local ice movement may have resulted in the transportation of an abnormally large quantity of debris into the main lake in the Columbia River valley. The massive silt layers associated with the varved silt may have been caused by such changes. They may be interpreted as drainage varves, similar to those which in some places have been found associated with desposits of normally varved glacial clays and so described and interpreted by Antevs (1922, p. 69) and others. The massive silt member of the three-part sequence may also be so interpreted since fluctuations in the level of the main lake and the formation and drainage of high-level lakes were both initiated by oscillatory movement of the ice margin and hence are related in time and basic cause.

Therefore, variations in the type of sedimentation in the Basal Sequence must have been dependent not only on the absolute depth of water in the main lake but also on the filling or sporadic drainage of temporary high-level lakes. Furthermore, repeated lithologies, especially the rhythmic repetition of the three-part sequence, must imply oscillations in the position of the ice margin just as do the concentric series of the more common looped recessional moraines.

The complete data and conclusions offered by Krynine in his study of the Basal Sequence silt sample seem to indicate that the sediment could have been derived from both the Okanogan region downstream to the northwest (chiefly by glacial transport) and from the San Poil and similar districts upstream to the northeast (chiefly by fluvial and lacustrine transport). This supports the conclusion reached on the basis of other field evidence. Incidentally, the presence in the Basal Sequence of fresh

euhedral crystals of calcic plagioclase, olivine, and hypersthene are believed to represent direct ash falls on the Coulee Dam locality during the accumulation of the aqueous sediments.

After the glacier had crossed the Columbia canyon and while it advanced south over the plateau, the Grand Coulee served as a spillway for the lake overflow, and the upper portion of the Basal Sequence was being deposited at the site of the dam. The succession consisting of (1) the till with conspicuous sand and gravel beds immediately above and below it, and (2) the Later Deposits (late-glacial Nespelem silt), which successively overlie the Basal Sequence west of the river clearly record the passage of ice over the site of the dam and the early phases of deglaciation.

During its maximum advance the ice blocked even the Grand Coulee and forced the overflow to still higher elevations in its passage across the plateau along the margins of the ice sheet and through spillways farther to the east (Flint, 1935, p. 189).

Early excavations near the west abutment of the dam revealed stratified sand and gravel dipping steeply west toward the valley wall. They rested on sheared and contorted varved silt and were unconformably capped by a layer of till. They are interpreted as high-level outwash which had slumped during a minor retreat of the ice front and which was subsequently overridden as the ice margin resumed its advance.

EAST SIDE OF RIVER

The deposits east of the river are less simple than those on the west, and the presence in them of deltaic deposits indicates an event in the later stages of deglaciation not heretofore recognized. As pointed out in the descriptive section, till grading up into foreset fluvial deposits rests unconformably on distorted varved silt of the Basal Sequence. Since the foresets dip downstream and are clearly deltaic, the materials must have been deposited when the Columbia River flowed into a lake at this point. Since their position in the Columbia canyon is several hundred feet lower than the floor of the nearby Grand Coulee spillway, the outlet of the lake at this time must have been downstream across the ice dam. The deltaic deposits unconformably overlie the varved silt of the Basal Sequence and therefore postdate the advance of the ice.

As explained later, the deltaic deposits must have formed after the deposition of sand, gravel, and Nespelem silt overlying the Till Zone west of the river, yet they occupy a position in the valley which should have been filled by this sand, gravel, and silt. This absence may be explained either by erosion or by nondeposition due to the presence of residual ice. Evidence seems to favor the latter, as will be indicated.

Deglaciation of an extensive downstream segment of the canyon would

TERRACED FILL UPSTREAM FROM GRAND COULEE DAM

View north across Columbia River (lower right) into Plum Basin, between Grand Coulee Dam and Keller Ferry, showing fill consisting chiefly of Nespelem silt. The highest continuous bench is the Nespelem silt terrace. (Photo by K. S. Brown, U. S. Bureau of Reclamation.)

FIGURE 1. DELTAIC SEQUENCE
Exposure in stream terrace (*cf.* Pl. 2, fig. 2) north of dam and east of river showing several courses of foreset structures dipping north (downstream).

FIGURE 2. SECTION IN HIGH, ICE-MARGINAL TERRACE
Exposure in Brett gravel pit showing till overlying stratified sand, silt, and pebbles. (Photo by K. S. Brown, U. S. Bureau of Reclamation.)

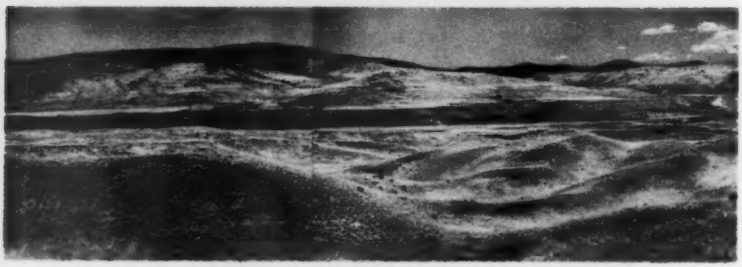


FIGURE 3. PITTED FACE OF THE NESPELEM SILT TERRACE
In Strahl Canyon, downstream from Grand Coulee Dam. View north, showing Nespelem silt terrace (high faint horizontal line) and conspicuous stream terrace (lower continuous line) on north bank of Columbia River.

STRATIGRAPHY AND MORPHOLOGY OF THE FILL

have been necessary to permit the return of normal fluvial conditions and the removal of these deposits. The deposition of the till which underlies and grades up into the deltaic material would then have required a readvance of the ice almost as extensive as the earlier advance. The continuity of the fill section west of the river, and the lack of evidence of a major readvance of the ice following deposition of the Nespelem silt, make the erosion hypothesis highly improbable.

The existence of residual ice masses in the Columbia canyon downstream from the mouth of the Grand Coulee at late stages in the deglaciation is indicated by ice-contact depressions in the face of the Nespelem silt terrace at several points, increasing in distinctness downstream. Flint (1935, p. 188) has pointed out that the absence of the Nespelem silt terrace in entrants in the valley wall near Grand Coulee Dam may indicate nondeposition due to the presence of residual ice. In addition, the existence of the deltaic deposits at low elevations in the valley seems to call for the continuous presence of residual ice until a late date in the process of deglaciation. The conditions leading to the formation of these deltaic deposits can best be explained by considering some of the processes active during the time immediately following the close of the Nespelem silt episode.

The final glacial event was the ultimate destruction of the ice dam and the disappearance of ice from the Columbia canyon. During the rebirth of the Columbia River, when the glacial lake was being restricted to successively lower reaches of the valley, it seems probable that three things were happening: (1) Erosion by the Columbia River of the recently exposed lake fill (Nespelem silt) along the upper reaches of the canyon; (2) transportation of the eroded debris downstream and its redeposition as deltaic material at the margin of the shrinking ice; (3) deposition of some of the debris as low-lying lacustrine sediments in the diminishing lake.

Strata possibly representing this ultimate low-level lacustrine phase are exposed along the banks of the Columbia River several miles downstream from Barry. However, the lithologic differentiation of deposits of this stage from similar materials belonging to a much earlier period in the ponding is difficult. Failure to identify them elsewhere may be due in part to their having been destroyed, since they must have been deposited along the axis of the canyon, which is the position now occupied by the river.

The zone of delta building that separated the normal Columbia River from the lake in its final low-level stages migrated downstream as the lake diminished. Although this zone was probably originally restricted in general to the axis of the canyon, now occupied by the river, marginal remnants of the deltaic deposits may have been preserved by lateral

shifting of the river channel during erosion to the present profile. It is believed that the foreset beds exposed east of the river are a remnant of this deltaic zone.

The gentle down-valley slope of the lake floor (the surface of the Nespelem silt terrace), and the absence of marked depressions in its surface, must have caused the transition zone between lake and river to be fairly broad and long. The deltaic beds deposited in this shallow-water transition zone would have been thin and difficult to identify. However, when the zone of delta building had migrated down-valley to the head of Grand Coulee it reached an area once covered by ice. Here the character of the lake floor must have been markedly different. The gradual wastage of residual ice probably had formed large depressions in the lake floor, and the irregular deposition of debris released directly from the ice front during the earliest stages of deglaciation must have further contributed to the irregularity of its topography. Furthermore the lake floor, though continuous upstream, must have graded downstream, within a short distance, into isolated constructional terraces that had been built between the canyon walls and ice in the central portion of the canyon.

Thus when the constantly shrinking lake had withdrawn from that portion of the smooth-floored canyon upstream from the head of the Grand Coulee, relatively deep water must still have existed in the canyon downstream from the head of the coulee. Therefore the conditions for the formation of thick deltaic deposits here were favorable.

A large quantity of coarse glacial debris deposited during oscillations of the ice front, and released by the still later wastage of residual ice, must have existed on the floor of the lake in this area. When the migrating zone of delta building reached the head of the Grand Coulee, deposition must then have taken place on a floor highly irregular in topography and varied in character of material. It seems improbable that the existing glacial debris could have escaped partial incorporation in the delta-building operation. Some of the glacial debris may also have slumped into the channel from unstable positions on the canyon walls. Some was certainly being released by the last vestiges of wasting ice. The process of delta building operating under such conditions must have resulted in an intricate arrangement and distribution of nonstratified coarse till-like debris and well-sorted fluvial material with intermediate gradations. As the delta-building process continued, the coarse, poorly sorted debris showing vague foresets was covered by better-sorted material carrying a higher content of river-borne gravel and showing foreset and associated stratification with greater clarity.

At several locations in the excavation, several courses of foresets resting one upon the other (Pl. 5, fig. 1) indicate that the surface of the lake in

which they formed may have been rising at times, temporarily reversing the general fall of lake level at this stage. Rise of the lake surface may record temporary readvance of the ice margin. That late readvance (or readvances) took place is supported by the evidence of ice shove now visible in the west bank of the river about a mile below the dam (Flint, 1935, p. 190). However, fluctuations in the surface of the lake during these final stages could have been due to damming caused by slides of drift from unstable positions on the valley walls.

The common occurrence of rhyolitic and andesitic rock types in the upper portions of the fluvial debris, together with their decreasing abundance at lower elevations and their almost total absence from the Basal Till, is believed to provide considerable support for the genesis of the delta.

According to the geologic map (Pardee, 1918, Pl. 1) of the country north of Grand Coulee Dam, considerable areas of Tertiary(?) rhyolitic and andesitic lavas occur. If a line is drawn due north from the site of the dam, the outcrop areas of these lavas lie almost entirely east of it, with only one very small area extending west of this line. Since the glacier is believed to have reached the vicinity of the dam by regional southeasterly flow, it is unlikely that fragments of these lavas could have been transported to the dam in quantity by glacier ice. This inference is supported by the fact that, not only in the till underlying the deltaic material, but in all other exposures of till studied in the vicinity of the dam, only two or three fragments of these lavas were found. On the other hand, the erosional debris from the areas of rhyolitic and andesitic flows which lie east of the N-S line would have been carried into the San Poil River and the upper reaches of the Columbia and would have been ultimately carried by the Columbia River past the site of Grand Coulee Dam.

RELATED FEATURES UPSTREAM AND DOWNSTREAM FROM THE DAM

SECTION NEAR KELLER FERRY

It is now known that the lake in which the Nespelem silt was accumulated extended from Grand Coulee Dam upstream uninterrupted only as far as the San Poil River, because the Columbia canyon was blocked for a time by the San Poil glacier lobe. On the basis of exposures existing in 1934, Flint (1937, p. 209) suggested that the San Poil glacier lobe might have approached the Columbia River but he could not find conclusive evidence of ice within 6 miles of the river at this point. However, grading on State Highway No. 4, in secs. 8, 17, and 20, T.28N., R.33E., on the south side of the river near Keller Ferry (at the extreme right in Pl. 4), exposed evidence which was examined in 1938. It consists of till inter-

calated within layers of the normal lacustrine fill of sand, silt, and clay, associated contortion of the lacustrine sediments, and a great accumulation of poorly size sorted rock fragments up to 3 feet in diameter, foreset west down the Columbia. The predominantly nonbasaltic lithology of the latter deposit indicates a source north rather than south of the Columbia River. The till lies at elevations between 1130 and 1170 feet; the rubbly accumulation continues up to 1550 feet. The stratigraphic relations of this sequence (although of course not the till and rubble themselves) are shown diagrammatically in an earlier publication (Flint, 1936, Pl. 6).

The San Poil glacier lobe therefore reached the south side of the Columbia canyon, although its terminal width was probably no more than 2 or 3 miles. The highest exposed till, indicating glacier ice at this locality, lies 500 feet below the top of the lake deposits, whereas the upper limit of the coarse ice-derived drift, only 100 feet below the top of the lake deposits, records ice near, but not immediately at, the site of the exposure. This seems to indicate that slight deglaciation had occurred while the last 500 feet of sediment was deposited in the lake. This parallels the relation exhibited near the dam, where the lake filling was not built up to its maximum height until deglaciation had progressed to a considerable degree.

SECTIONS BETWEEN THE DAM AND NESPELEM RIVER

Along the Grand Coulee Dam-Nespelem road northwest of Peter Dan Creek, at elevations between 1100 and 1250 feet, there are several exposures in a thick sequence of silt and clay, which exhibit abundant intercalated stones probably ice-rafted, contorted lamination, and thin layers of till. Along the same road near the intersection of the side road to Barry Ferry, at elevations of 1450 to 1500 feet, a 4-foot till layer and a rubbly bouldery gravel are intercalated within the silt. The top of the continuous section in this vicinity, at an elevation of 1700 feet or more, consists of coarse sand and gravel with current bedding indicating flow toward the southeast. Probably this represents fluvial deposits derived through the Nespelem valley at the climax of lacustrine filling.

Excellent sections are exposed in two large draws entering the Columbia River from the east in T.30N., R. 30E. One draw, in sec. 26, exposes 400 feet of little-disturbed laminated silt, resembling the silt above the till at the dam, with, near the top, a conspicuous zone of till and boulders (including one 30 feet in diameter). The other draw, in sec. 23, exposes 500 feet of mostly little-disturbed laminated silt and clay (Pl. 6, fig. 1), of which the upper 200 feet is locally contorted and includes several zones of till, boulder beds, and a large number of single boulders 10 to 30 feet in diameter.

FIGURE 1. LATER DEPOSITS
Section consisting chiefly of laminated silt, with some interbedded till, exposed on Columbia River near the mouth of Nespelem River. View north. (Photo by 116th Photo Section, Washington National Guard.)

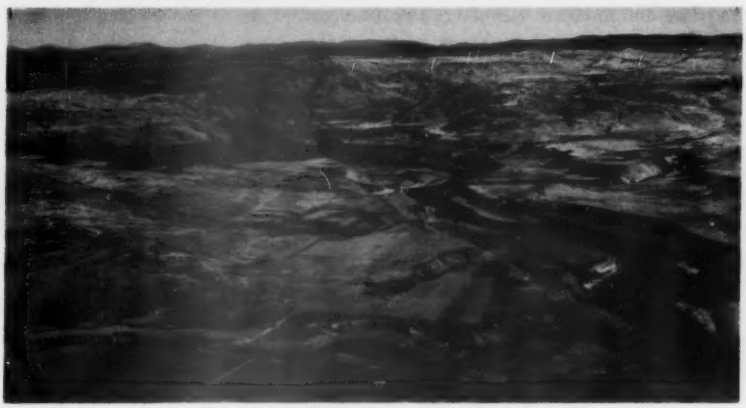


FIGURE 2. DISSECTED FILL
View northeast up Columbia River from near Omak Lake Trench, showing Nespelem silt terrace (discontinuous high remnants) and the conspicuous stream terrace (broad surface in foreground and on right). (Photo by 116th Photo Section, Washington National Guard.)

GLACIAL FILL DOWNSTREAM FROM GRAND COULEE DAM

All these sections clearly record a number of minor ice advances into a lake in which thick fine-grained sediments were accumulating, with crumpling by overriding, deposition of layers of till and boulders, and dropping of large erratics onto the lake floor. The sequence represented by these sections is analogous in origin with the Till Zone at the dam 5 to 10 miles upstream. However, it is believed not to be correlated in time with the Till Zone at the dam, because silt that is conspicuously undisturbed by ice lies between these sections and those at the dam. It is probable that the till downstream was made after deglaciation of the site of the dam and after partial filling of the valley axis with silt, but while the lake still discharged through Grand Coulee and while the active ice terminus was shrinking back to positions several miles north of Grand Coulee Dam.

In the many exposures of silt examined through a long distance down the Columbia River below the Nespelem River, neither till nor erratics were seen interbedded with the silt. In consequence, it is possible that by the time deglaciation had proceeded this far, the ice remaining in the canyon had ceased to be active, and was gradually wasting out and permitting the space it had occupied to be replaced by normal lacustrine deposits.

COLUMBIA CANYON BETWEEN NESPELEM RIVER AND OKANOGAN RIVER

The Nespelem silt terrace is a prominent feature in the Columbia canyon as far downstream as the Omak Lake Trench, but west of that place it is not present. The terrace consists of silt with considerable sand and gravel, especially north of the river. The coarser sediments appear to be chiefly fluvial contributions from ice to the north, via small northern tributaries to the Columbia River. The elevation of the upper surface of this terrace rises from about 1700 feet near Grand Coulee Dam to about 1800 feet near the Omak Lake Trench. This, together with the fact that this surface carries no veneer of rounded gravel like the veneers on the lower terraces, is believed to indicate that in this part of the canyon drainage was moving up the canyon toward the Grand Coulee spillway outlet up to the close of the episode in which the sediment of this terrace was accumulated.

Below the Nespelem silt terrace are discontinuous remnants of terraces cut from the fill by the Columbia River (Pl. 6, fig. 2). These are nonpaired, occur down to the present profile of the river, and are capped by thin covers of rounded gravel, the peculiar lithology of which has been indicated. The terrace which is generally broadest, most nearly continuous, and most conspicuous, lies at an elevation of 1300 to 1400 feet (Pardee, 1918, p. 15) in this part of the canyon. Unlike the Nespelem

silt terrace, this stream terrace is present (though not continuously so) down the Columbia River beyond the Omak Lake Trench. Downstream as well as upstream it is cut from a fill that is chiefly laminated silt. It appears to be continuous with a conspicuous terrace still farther downstream below the mouth of the Okanogan (Flint, 1935, p. 188), the whole system apparently recording a temporary base level that interrupted for a time the otherwise steady progress of dissection of the fill.

The Nespelem silt terrace is prominently knolled and kettled (Pl. 5, fig. 3) at several places near its downstream terminus. These are:

S 1/2 sec. 20, T.31N., R.29E., in Hopkins Canyon, a north tributary.
Sec. 19, T.30N., R.29E., in Strahl Canyon, a south tributary.
Sec. 31, T.31N., R.28E., at junction of the Omak Lake Trench.
Secs. 15, 16, and 21, T.30N., R.28E., opposite junction of the Omak Lake Trench.

Not only does the Nespelem silt terrace exhibit these ice-contact features, but the broad conspicuous stream terrace is knolled and pitted in sec. 3, T.30N., R.28E., and in sec. 30, T.31N., R.28E., both localities being in the Omak Lake Trench near its junction with the Columbia. These features apparently indicate that glacier ice still blocked the Columbia canyon west of the Omak Lake Trench when the sedimentation that built the Nespelem silt terrace fill was brought to an end by abandonment of the Grand Coulee spillway and resumption by the Columbia River drainage of its former course. They seem to indicate further that lacustrine deposition continued to fill the places along the axis of the canyon that were left empty by ice wastage, even after the Grand Coulee was abandoned, and that renewed drainage westward over the old route took place, for a time at least, over buried ice, which did not waste away until after the river had abandoned the profile recorded by the conspicuous stream terrace and which had cut down to the profiles recorded by still lower terraces. Active ice was present not far north, however, because a valley train with its ice-contact head at the south end of Omak Lake descends south through the trench and joins the Columbia on a profile that approximates that of the broad conspicuous stream terrace. The valley train has been incised by meltwater from the same source to form terraces that reach the Columbia River on still lower profiles. This shows that ice was present in the Omak Lake Trench down through very late stages in the dissection of the fill in the Columbia canyon by the Columbia River.

DATE OF GLACIAL DEPOSITS

The glacial deposits in the Grand Coulee Dam district are clearly the product of the latest glaciation that affected the district. The date of this glaciation has been rather generally assigned to the Wisconsin stage (*e.g.*,

Flint, 1937, p. 224). No direct evidence of an earlier glaciation was seen in the vicinity of the dam during the present investigation.

The (upper) Grand Coulee channel and intake must have existed prior to the period of maximum ice advance in the latest glaciation, because it is partially filled with glacial drift and Nespelem silt which were deposited during deglaciation. Furthermore, indirect evidence not yet published leads the authors to the belief that the erosion of the Grand Coulee channel was complete (or nearly so) prior to the arrival of glacier ice in this district during the latest glaciation. Therefore earlier glaciation is implied, if, as is commonly accepted, the erosion of the spillway was accomplished by the overflow of glacially ponded water from the Columbia canyon.

GENERAL CONCLUSIONS

(1) Sections of the fill in the Columbia canyon, together with the stratigraphy and morphology of related features downstream, record the expansion of the Okanogan glacier lobe to its maximum, and its later shrinkage and disappearance, during the last glaciation.

(2) The Basal Sequence at the dam records: (a) In its lower alternately fluvial and lacustrine part oscillatory damming of the Columbia River downstream when the ice first encroached upon it to form a shallow lake that drained west around the ice margin; (b) in its upper dominantly lacustrine part, the final and complete damming of the Columbia River and the gradual deepening of the lake, with outflow via the Grand Coulee route.

(3) The Till Zone, overlying the Basal Sequence, records the arrival of ice at Grand Coulee Dam with oscillatory fluctuations of the ice margin during the climax of glaciation.

(4) The San Poil glacier lobe entered the lake-filled Columbia canyon at Keller Ferry, 15 miles upstream from Grand Coulee Dam, and contributed sediment (part of the Nespelem silt) to the lake. The maximum of this lobe appears to have been approximately contemporaneous with the maximum of the neighboring Okanogan glacier lobe.

(5) The later deposits, stratigraphically (but not everywhere vertically) overlying the Till Zone, record progressive deglaciation accompanied by minor re-expansions of the ice and by transition from lacustrine to fluvial conditions. The till sections downstream are in part the time equivalent of (a) the Later Deposits at Grand Coulee Dam, and (b) the upper part of the Nespelem silt upstream from the dam.

(6) While residual ice still partly filled the Columbia canyon at and west of the Omak Lake Trench, drainage from the lake began to escape over and around it via the pre-lake route, abandoning the Grand Coulee

route. This event was contemporaneous with and slightly later than (a) the first of the Later Deposits at Grand Coulee Dam, and (b) stream terraces cut from the lake filling upstream from the dam at least as far as the San Poil River.

(7) Late-glacial and postglacial grading of the long profile of Columbia River proceeded by (a) filling of residual pools in ice-block depressions and in depressions dammed by ice (here the deltaic deposits at Grand Coulee Dam), and (b) contemporaneous trenching of slightly earlier lake filling elsewhere (here the stream terraces both up- and down-stream). Grading down to the present river profile was accompanied by at least one marked delay owing to a temporary base level downstream recorded by a persistent stream terrace more conspicuous than the others.

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FIGURE 1. FROM THE EAST, AUGUST 1911
Photo by T. W. Stanton, U. S. Geological Survey.

FIGURE 2. FROM THE EAST, AUGUST 29, 1935
The glacier consists of two separate ice bodies in contrast to the condition depicted in Figure 1. The north moraine lies at the extreme right edge of the lower ice mass.

GRINNELL GLACIER

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GRINNELL GLACIER, GLACIER NATIONAL PARK, MONTANA

BY GEORGE R. GIBSON AND JAMES L. DYSON

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ABSTRACT

Stratification due to the annual accumulation of snow is well exhibited in Grinnell Glacier. Layers of snow in the narrow firn zone at the head of the glacier are wedge-shaped and have a steep initial tilt toward the glacier front. Weight of overlying layers forces the earlier-formed layers to rotate so that after travelling a relatively short distance they assume a backward dip which gradually increases until, approximately halfway between cirque wall and glacier front, a maximum of 45 degrees is attained.

Detailed study of the position of ice strata in various parts of the glacier has yielded information on the nature of the ice motion, the age of the ice, and its probable thickness.

Available records dating back to 1897 and direct observation by the authors over a 4-year period indicate that Grinnell Glacier has been reduced to about half the size it was in 1900, and that the recession during recent years has been most rapid.

INTRODUCTION

Most of the data for the preparation of this paper were secured during the mapping of Grinnell Glacier in the latter part of August 1937 and from observations made at intervals during the summer months of 1935, 1936, and 1938.

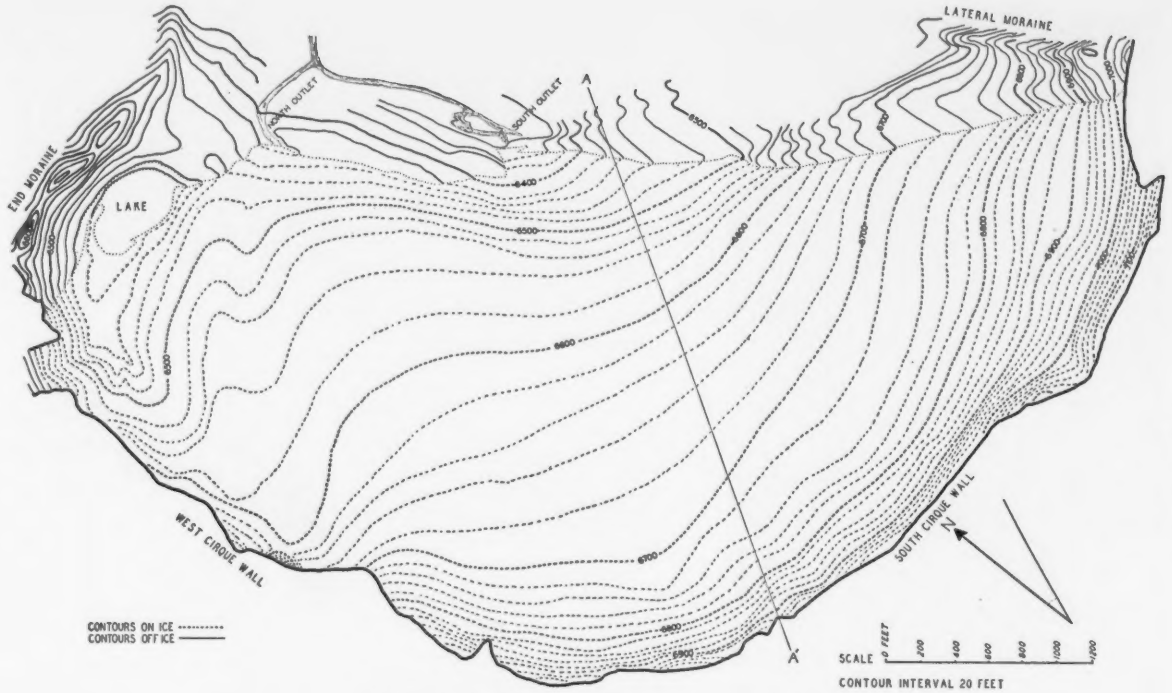

FIGURE 1.—*Topographic map of Grinnell Glacier*

Elevations based upon datum determined by aneroid. (Used with permission of The Director, National Park Service, Washington, D. C.)
For section A-A', *see* Figure 4.

The writers felt that a detailed study of a cirque glacier could be expected to yield interesting results. Large, rather than small, glaciers have usually been most carefully examined, and attention has been directed mainly to their broader features. Details which might provide the keys to understanding of important glacial phenomena have therefore often been neglected. Further, a small cirque glacier in an extreme state of decadence, as is the case with Grinnell, exhibits evidence of its structure and displays other glacial characteristics very clearly. Such a small cirque glacier has motion, deposits morainic material, and performs other typical glacier functions. If it is also receding, the results of these functions tend to be progressively exposed. Additional striae and moraines are being uncovered yearly, and all parts of the glacier are made accessible because of the nearly complete disappearance of snow from its surface in the late summer.

The authors wish to acknowledge their indebtedness to the National Park Service for the opportunity to carry on the work, and to Professor L. M. Gould and Professor O. D. vonEngeln for suggestions during the preparation of the manuscript.

LOCATION AND GENERAL DESCRIPTION

Grinnell Glacier, one of the larger ice masses in Glacier National Park, is situated on the east side of the Continental Divide at the head of a tributary of the Swiftcurrent Valley. Its melt water drains by way of Swiftcurrent Creek and the St. Mary, Saskatchewan, and Nelson rivers into Hudson Bay.

Grinnell is representative of most glaciers in the region in that it lies in a conspicuous steep-walled cirque, the floor of which is a shelf situated about 1300 feet above the main part of the valley which was formerly occupied by a large valley glacier. The Chief Mountain, Montana, and Glacier National Park, Montana, topographic sheets of the United States Geological Survey show Grinnell Glacier (latitude 48° 45′ north) as mapped in 1900. At that time and until sometime between 1911 and 1929 the glacier consisted of two parts, a very steep upper part lying on a bench in the western part of the cirque and a lower larger part—the main mass of the glacier. The upper part was connected to the lower by a narrow, much crevassed tongue of ice (Pl. 2, fig. 3). The separation of the glacier, prior to 1929, has given rise to two independent ice masses. It is the lower and larger of these that is herein considered and is referred to as Grinnell Glacier.

The elevation of the lowest part of Grinnell Glacier in August 1937 was approximately 6300 feet as determined by aneroid, and the difference in elevation between this point and the highest part of the ice was 840 feet

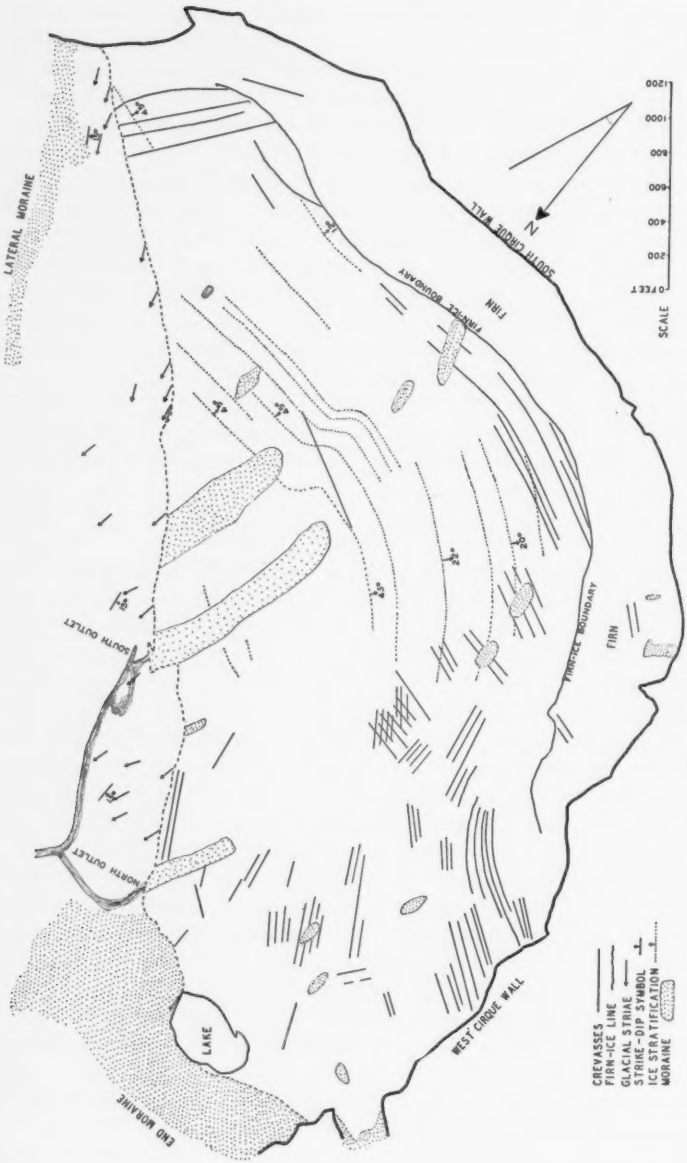

FIGURE 2.—*Grinnell Glacier in August 1937*

(By permission of The Director, National Park Service, Washington, D. C.)

(Fig. 1). The surface area at that time was about three-quarters of a square mile.

The longest dimension of the glacier (northwest-southeast) was 6800 feet, only slightly less than the same dimension as recorded on the U. S. Geological Survey maps. This figure gives the approximate maximum extent of the glacier from side to side. At a right angle to this direction the ice mass had a length of 3200 feet, which is considerably less than the same dimension on the United States Geological Survey maps. This value is also smaller than that recorded by Alden (1914, p. 19) from his observations in 1911 when the glacier had a length of about a mile (northeast-southwest).

On its west and south sides Grinnell Glacier is bounded by the vertical walls of the cirque. The east border of the ice for the most part is thin and rests on a smoothed and striated bedrock surface. This edge of the glacier has been retreating more rapidly than the north border of the ice. An abandoned lateral moraine roughly parallels the east edge of the glacier. This morainic embankment was separated from the ice in August 1937 by a distance varying from approximately 200 feet to a quarter of a mile. The moraine at the north side of the glacier is much higher and more conspicuous. Between the north edge of the ice and this moraine there is a small lake immediately east and west of which (Fig. 1) the ice front rests against the base of the moraine. This lake, because of the presence of snowdrifts, does not usually appear until late June. In August 1937 several conspicuous blocks of ice broken from the glacier appeared as icebergs in the lake (Pl. 2, fig. 2). At its eastern end the bottom of the lake is a grooved and striated bedrock surface.

The ice in places is prominently crevassed and relatively free of debris except for thin morainal bands along the east border and isolated "debris islands" scattered over the surface (Fig. 2).

ICE STRATIFICATION

If the glacier is viewed from a point several hundred feet above its surface and from the east or west sides, parallel lines will be noted sweeping across the ice in a series of graceful curves more or less conforming to the configuration of the cirque wall (Pl. 3, fig. 1). They are most prominent in front of the snow-covered area which bounds the south wall of the cirque.

At places these zones or bands are about 2 to 3 inches in width and are composed of ice mixed with dust, organic material, and fine particles of rock. Because of the greater absorbtion of the sun's heat by this material, the bands are made more conspicuous by being depressed slightly below the surface of the ice. At other places where the ice is charged

with more debris, these zones may be several feet wide. They are from 6 to 15 feet apart on the surface and have a dip of 12 degrees to the south at the firn-ice boundary, but gradually become steeper until at some points 1100 feet from the border of the firn they attain a maximum dip of 45 degrees to the south. Thus the thickness of the ice between the zones is 4 to 6 feet. Because of minor irregularities of the ice surface these dips can be readily noted (Pl. 3, fig. 2). In crevasses the dips of these zones can be measured to depths of 40 feet below the surface. At each site examined the dip was found to be constant to that depth.

The banding of glacial ice has been often discussed in the literature, and there seems to be general agreement among observers in regard to its causes. The type of banding found on Grinnell Glacier, for the reasons cited above, is that which most authors describe as the boundary planes or zones between successive snowfalls. Agassiz (1840, p. 40) mentioned it, and Russell (1897, p. 43), in discussing the glaciers of the Sierra Nevada, called the phenomenon "dirt bands" and said that these were to be observed on nearly all the glaciers of that region. He described them as discoloring the ice in well-defined strata dipping into the glacier at a low angle with the surface. Sherzer (1907, p. 118-119) described several types of banding in the glaciers of the Canadian Rockies and attributed one type to stratification resulting from the separation plane between successive annual accumulations of snow. The stratification in Grinnell Glacier corresponds to these descriptions. Sherzer designated the surface manifestation of the stratification "dirt zones" in order to make clear the distinction between them and the "dirt bands" of Forbes, which are due to dirt accumulating in open crevasses that later close as they move toward the glacier front. Seligman (1936, p. 273) said:

> "A form of stratification is particularly well seen in the firn snow region of glaciers, especially where crevasses have laid bare enormous sections. There is little doubt in the minds of modern glacialogists that these strata tally with the seasonal precipitation, although it seems probable that during the latter part of the firnification process the individual falls of snow become absorbed. When stratification is apparent in true ice this is still more certain."

The location, size, and shape of the snowfield area adds support to the interpretation that this banding on Grinnell Glacier can be attributed to seasonal falls of snow, with each season contributing a stratum. On Figure 2 the snow-ice boundary can be noted extending from the east side of the glacier westward to the point where the cirque wall turns due north. The shape of the firn field is that of a crescent slightly over a mile in length and 600 feet wide at the center. Its location is at the head of the glacier resting snugly against the cirque wall. The total area occupied by the firn field is approximately 0.06 square mile or about 8 per cent of the surface of the glacier.

Several factors govern the size and shape of the firn area. Most of the day the back portion of the glacier is protected from the sun's rays by a high vertical cliff, preserving the snow there from much melting, while the portion of the glacier not so protected loses all its winter accumulation of snow during the summer. The cirque wall lessens the velocity of the wind in its immediate vicinity also. Farther out on the ice, the wind is distinctly effective in evaporating snow, especially in summer. The prevailing west wind sweeps over the cirque basin in the winter, causing a whirlpool of air which drifts the snow back against the cirque wall, piling it to a much greater thickness there than elsewhere on the glacier. The authors studied these currents from the top of one of the cliffs above an abandoned cirque late in August. A map was snatched by the wind and carried out over the basin 2000 feet below. As the map moved toward the open end of the cirque it gradually settled and when a certain level was reached it came back toward the foot of the cirque wall and rose slightly, then settled. This type of wind action and the fact that the snow melts from the rest of the surface would be responsible for the firn field being the steepest part of the glacier with the greatest declivity next to the bergschrund. Therefore, the protection offered by the cliff at the rear of the cirque and the drifting action of the snow by the wind would account for the location, shape, and size of the firn field.

EXPLANATION OF DIP OF ICE STRATA

In observing the stratification lines, it was noted that the dip was approximately 12 degrees to the south at the firn line and gradually became steeper until about halfway to the glacier front at places it reached a maximum of 45 degrees to the south. A similar observation was made on the Victoria Glacier by Sherzer (1907, p. 38-39). He says:

> "These strata all dip toward the regions of accumulation, directly beneath the front of the hanging glacier. In the lower part of the glacier this dip averages 22 degrees, ranging from 12 degrees to 26 degrees, while farther up-stream the dip is more gentle, only 5 degrees to 10 degrees, as well seen in the crevasse walls."

Why should the dip increase as the ice moves toward the front? An answer to this question might tell us something concerning the movement of ice in the vicinity of the head of the glacier.

The reservoir of the glacier is the firn field which lies against the cirque wall where the ice receives protection from the sun and wind. It is crescent-shaped, slightly over a mile in length, and about 600 feet wide. It probably is at least 40 feet thick at the cirque wall and tapers to a thin edge toward the north. Away from the firn field the ice strata range between 4 and 6 feet in thickness, thus requiring 40 to 60 feet of snow for the formation of each layer. This amount of snow seems large

in a region where the annual precipitation is approximately 30 inches. However, drifting apparently accounts for it.

Each winter a new layer of snow is added, dipping north since that is the direction of the slope. In summer, dust, particles of rock, and organic material are deposited on the surface by the wind, thus making

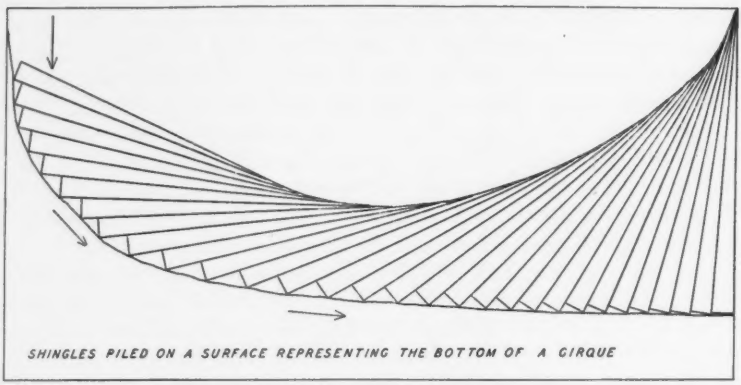

FIGURE 3.—*Diagram of shingle model*
Showing movement of ice strata in Grinnell Glacier.

a visible bounding plane between each season's accumulation. Each season's snow is compacted into a wedge-shaped mass of granular snow, thickest at the cirque wall. By its own weight and that of the later layers this wedge moves downward.

The bedrock floor upon which the glacier rests may be either a nearly horizontal surface or a distinct basin. Both these types of floors are exhibited by other cirque basins in the region. Assuming either, the ice layers as they move down the cirque wall and then over the floor would have to tip backward, and the dip of the stratification planes would become steeper with increasing distance from the cirque wall.

A model (Fig. 3) was made which illustrated this rotating movement at the rear part of Grinnell Glacier. A number of shingles were used in experimentation for the development of the model. Each shingle, because of its wedge shape, represented one season's accumulation of snow. The profile of the cirque floor was that of a vertical backwall joined to a nearly horizontal floor by an arc. If a complete basin with the floor a surface concave upward were used, approximately the same rotating results would be attained.

The shingles, with the thick end at the backwall, were placed one on top of the other and pushed downward. The curvature of the lower

part of the backwall forced the shingles to rotate, taking a higher dip backward as each shingle moved downward and forward. This type of rotation would continue until the more level surface of the cirque floor is reached and would probably be inhibited where the downward pushing of the later layers ceased and where the horizontal pushing became the more important. In the case of a basin-shaped floor the layers would rotate slightly until the lip of the basin is reached. If this model indicates the ice motion near the head of the glacier, the thin part of each wedge is removed by melting before it has travelled far and only the ice which was originally next to the cirque wall, or which composes the thick end of the wedge, reaches the glacier front. The strata starting out inclined toward the front rotate and take a backward dip while passing over the arc connecting the cirque wall and the floor.

There is other evidence that the ice probably moves in this way. Piles of rock debris are found on the ice from a short distance in front of the firn line to the ice front. The material forming these piles came from the cirque wall at some distance above the glacier, as can be noted by the scars left by rock avalanches. Because of the steepness of the firn slope they probably came to rest near the firn line and were buried by the next season's snowfall. However, they soon reappear because of melting and ride on the surface to the glacier front. Even if a pile of rock were buried in the middle of the firn it would soon reappear at the surface; this seems to be the case because some of the largest piles are near the snow-ice line.

The fact that the stratification planes have a constant dip to a depth of 40 feet, as noted in crevasses, would indicate little differential movement across the planes. Most of the movement must have been parallel to the bedding planes. In this respect the glacier moves in a more or less rigid manner with approximately the downstream half moving as a unit. Plastic flow apparently takes place in a very thin zone along the base of the glacier, as indicated by the small amount or distortion in ice strata at places along the ice front and by the sinuous courses of striae adjacent to Grinnell and other ice masses of the Park (Demorest, 1938, p. 700-725).

With the aid of a hand lens on Figure 1 of Plate 3 a count of the stratification lines was made. From the south cirque wall to a point on the east front about 300 feet east of the easternmost morainic band (Fig. 2), a distance of 1800 feet, approximately 60 stratification lines were counted. If these lines are the surface expression of boundary planes between successive seasonal accumulations of ice, the time necessary for the ice in the eastern part of the glacier to travel from the cirque wall to the ice front is approximately 60 years. Thus the average rate

of movement for the last 60 years in that part of the glacier is about 30 feet per year.

FEATURES OF ICE MOTION

The direction of striae on bedrock adjacent to the ice, the trend of morainic bands on the surface of the glacier (Fig. 2), and the trend of ice stratification bands indicate that the motion of Grinnell Glacier is not from the cirque wall directly out to the ice front. Debris within and on the ice along the south cirque wall (Fig. 2) in its travel to the terminus follows a course which describes an arc convex to the west. Apparently the ice moves out from the south cirque wall and parallel to the west wall. Thus, based upon the orientation of morainic bands, striae, and ice stratification, it appears that the maximum horizontal distance ice within the glacier can move in its course from cirque wall to ice front is approximately seven-eighths of a mile. This distance then is the true length of Grinnell Glacier.

Notwithstanding its small size and apparent thinness, this glacier thus exhibits flow phenomena not unlike those of a long valley glacier which may follow a sinuous course from cirque head to terminus. Striae near the east border of the ice indicate that there was an even greater curvature in the lines of glacial motion 10 or 15 years ago, as striae, with increasing distance from the glacier, trend more toward the northeast. Ice originating in the firn zone along the south cirque wall moves in a direction somewhat west of north. After travelling a mile this ice must have been moving in a direction nearly 20 degrees east of north, about 30 degrees from its initial direction of motion.

The character of the material in the moraines themselves gives a further clue to the direction of motion of the glacier. That portion of the north moraine adjacent to the present north margin of the ice is a typical terminal moraine. In addition to much angular material, it includes numerous striated and glacially modelled boulders and much rock flour which acts as mortar binding the coarser materials of the moraine together. Thus, the inside or ice-contact slope of this accumulation is actually steeper than the angle of repose of the material. So firmly cemented is it that erosion and sliding of its surface have been negligible since withdrawal of the ice which began sometime prior to 1900. This surface of the moraine is marked by several small closely spaced ridges which run from the base of the moraine to its crest in the same direction that the former ice moved. This feature is much more pronounced on the ice-contact slopes of other moraines in the Park, particularly those adjacent to Sexton, Clements, and Sperry glaciers. On the Sperry moraine the ridges are so numerous that the inside surface of the moraine has a fluted or furrowed appearance. In addition to representing the former

direction of ice motion these small ridges indicate that the ice moved up over the inside surface of the moraine. Under these circumstances considerable material was plastered onto the surface of the moraine directly from the basal portion of the ice.

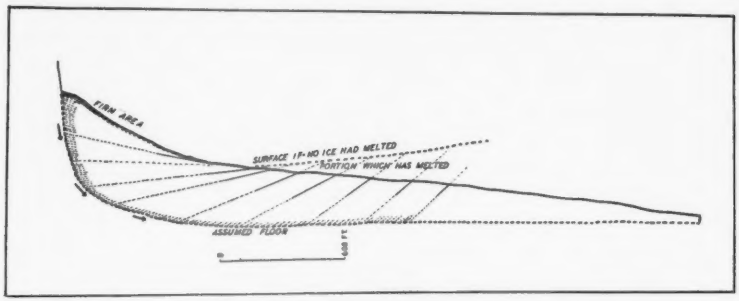

FIGURE 4.—*Idealized cross section of Grinnell Glacier*
Along the line A–A′ on Figure 1.

The east moraine, in contrast to the north, is lateral. As material within and on the glacier was carried parallel to this moraine it is considerably smaller, in spite of the long stand of the ice against it, than the terminal moraine at the north ice front. Its contents are mainly angular blocks of rock, many of which have come to rest in the moraine after sliding across the ice surface. Striated boulders are extremely rare in this accumulation, and rock flour is not present.

Present erosional activity of Grinnell Glacier is perhaps indicated best by the milkiness of its discharged melt water. Grinnell Lake, about a mile downstream from the glacier, acts as the first settling basin for the rock flour. Despite this half-mile wide expanse of still water enough rock flour persists to the outlet to give that stream a noticeably milky appearance. It is not until the water passes through Josephine Lake and then Swiftcurrent Lake, 5 miles from the glacier, that it becomes a clear stream.

The water in the small lake at the north end of the glacier and in the north outlet (Fig. 1), which apparently drains the lake, is very milky, whereas water of the south outlet is quite clear. It appears that the source of rock flour is the north and northwest portions of the ice which drain through the north outlet. The south outlet receives melt water from the southeast section of the glacier. There probably is a high area or divide on the cirque floor under the glacier between these two drainage areas. This is indicated by the trend of the ice stratification lines. When traced from the east edge of the ice they trend in an east-

west direction. East of the center of the glacier they turn southwest for 200 feet and then resume the east-west trend. This flexure might be caused by a ridge of rock underneath the ice; in fact, a ridge of rock which lines up with the strike of this flexure can be noted passing under the ice.

The fact that the ice of the southeast portion of the glacier is too thin to do significant erosive work seems inescapable. If such is the case, ice of the northwest part may be lying in a basin such as many cirque glaciers have produced in other sections of the Lewis and Livingston ranges. This basin might give the ice the necessary thickness required for erosion and the production of rock flour.

Another, but less likely, possibility is that the glacier has receded to the extent that, with the exception of the relatively small headward portion, it no longer is an agent of corrasion; most of the rock flour in the north outlet stream is being furnished by morainic material, which does contain much finely powdered rock, adjacent to and underlying the northwest section of the glacier. This material was produced when the ice was 100 feet or more thicker than at present. Such an inference entails the conclusion that the thickness of the ice in Grinnell Glacier, with the possible exception of that in the narrow firn zone, probably does not exceed 100 feet. Such a deduction is based upon the work of Demorest (1938, p. 725) at Clements Glacier and upon observations by the authors adjacent to several glaciers similar to Grinnell in the Lewis Range, which show that ice with a thickness no greater than approximately 150 feet is quite capable of performing marked corrasion on rock surfaces of sandstone, limestone, and shale.

RECESSION OF GLACIER

Since about 1900 practically all the ice masses in Glacier National Park have been undergoing rapid recession. That there was a period prior to 1900 during which these glaciers were nourished sufficiently to maintain their size is indicated by the conspicuous, now isolated moraines which encircle the sides and fronts of all the glaciers. In some cases these moraines attain heights of nearly 200 feet. That they represent a comparatively long stand by the ice fronts is indicated by their size; in addition, as observed adjacent to the east lobe of Blackfoot Glacier, the stand was of sufficient duration to enable streams marginal to the ice to entrench themselves in bedrock transversely to the slope.

Historical records of scientific value for the glaciers of Glacier National Park do not extend back much beyond 1900, so that the only means of estimating the latest approximate date when the glaciers completely filled the areas behind the now isolated moraines, or the duration

FIGURE 1. ICE RESTING AGAINST NORTH MORAINE, AUGUST 1911
Photo by T. W. Stanton, U. S. Geological Survey.

FIGURE 2. EAST FRONT OF GRINNELL GLACIER
North moraine in left foreground. Dashed line shows approximate boundary of glacier in August 1911. Photo taken August 29, 1937.

FIGURE 3. CONNECTION BETWEEN UPPER AND LOWER PARTS OF GRINNELL GLACIER IN AUGUST 1911
North moraine in foreground. Photo by T. W. Stanton, U. S. Geological Survey.

FIGURE 1. EASTERN HALF OF GRINNELL GLACIER
Showing stratification lines and firn-ice boundary.

FIGURE 2. LOOKING NORTHWEST FROM CENTER OF GLACIER
Showing dip of contact planes between ice strata. Photo taken August 30, 1937.

GRINNELL GLACIER

of this stage, may be obtained only from examination of the moraines themselves and from studies of the recent rate of recession. It seems certain, however, that 50 years ago these glaciers had the maximum size allowed by the position of the moraines which surround them today. That all glaciers of the area were affected in the same manner is indicated by the fact that isolated moraines occur around all the glaciers. They are particularly striking at Sperry, Blackfoot, Swiftcurrent, Sexton, Harrison, Clements, and others. Recession of all glaciers since beginning of withdrawal from the moraines has been more or less uniform, although a few which lie in well-protected sites have exhibited noticeably less rapid shrinkage than others.

Photographs [1] taken by E. R. Shepard in August 1897 show Grinnell Glacier resting high against the sides of the moraine which is now isolated. Comparison with photographs (Pl. 2, fig. 1; Pl. 1, fig. 1) by Stanton in August 1911 indicates that the glacier, although somewhat smaller in size, had not undergone an appreciable amount of recession during the previous 14 years.

According to Alden (1914, p. 19), in 1911 the morainal embankment closely bordering the east and north margins of the ice reached heights of 100 feet. The north moraine is the higher of the two, and in August 1937 its crest in places was nearly 200 feet above the surface of the ice (Fig. 1). Alden's statement and comparison of 1937 photographs (Pl. 2, fig. 2) with pictures taken in 1911 (Pl. 2, fig. 1) indicate that the surface of the glacier at the northern terminus has been lowered approximately 80 feet since 1911, thus exposing the lower part of the moraine against which the ice front rested in 1911. This shrinkage of the glacier caused the crest of the moraine to be about 80 feet higher above the ice than when Alden examined it in 1911.

The apparent increase in the rate of recession since 1911 is in part due to the fact that as the ice becomes thinner, particularly at its border, smaller amounts of surface lowering will cause greater recession of the edge of the glacier. A lowering of the ice surface by 80 feet since 1911 along the north margin of the glacier has resulted in only a small amount of recession of the edge of the ice because of its relatively great thickness. Thus the northern edge is not now far removed from the positions it had in 1897 and 1911, whereas the east border of the glacier, which in the past has been much thinner than the north, has shown a large amount of recession. In spite of its greater thickness in the past, the northern edge of the ice has been lowered until it is now very thin. Thus continued wastage of the ice will probably cause a very rapid withdrawal of the

[1] These photographs are now in the possession of Mr. A. L. Sperry of Owatonna, Minnesota.

north front which will, quite probably, correspond to the rate of recession of the east border.

A similar condition has apparently been attained by Agassiz Glacier which is the largest single ice mass in the Park. As reported by Ruhle (1937, p. 298), Park Naturalist of Glacier National Park, ablation prior to 1935 had reduced the ice tongue to a thin sheet, thus probably accounting for the unusually large amount of recession (465 feet) during the 12 months immediately preceding September 1936.

It is reasonable to believe that the west lobe of Blackfoot Glacier, which has likewise been reduced to a very thin tongue (Dyson, 1937, p. 552), will show an increased average yearly rate of recession in the future, provided climatic conditions of the past 10 years continue.

Since 1932 the yearly amount of recession of Grinnell Glacier has been measured each fall by the Park Naturalist. The site used for this work is the short ice tongue at the south outflow stream (Fig. 1) near the middle of the east front of the glacier. The retreat of this part of the ice margin is indicative of the distance of withdrawal all along the east front. During the 4-year period, 1934 to 1937 inclusive, this ice tongue receded a total distance of 165 feet or a mean yearly average of 41.2 feet.

If the ice front remained stationary for a period of years, piles of debris would be formed along the margin of the glacier at the ends of the morainic bands which are so conspicuous along the east border of the glacier (Fig. 2). However, as morainic material is no more abundant on the bedrock surface adjacent to the east edge of the ice than it is on nearby portions of the glacier, the yearly amount of recession (approximately 41 feet in recent years) since the glacier retreated from the east moraine has been as great as, or greater than, the distance the ice moves each year in the frontal part of the glacier. This phenomenon does not apply to the north margin of the ice as there the glacier rests upon a comparatively thick morainic deposit.

From August 26 to August 30, 1920, Alden (1923, p. 268) found by crude measurements that the maximum movement of the ice near the northeast margin (approximately the center of the glacier front) was slightly in excess of an inch per day. If this rate is maintained throughout the entire year the amount of movement would be approximately 35 to 40 feet. This rate corresponds rather closely with the figure derived by the authors from a count of the seasonal bands. However, comparison of the two figures indicates that the rate of motion is somewhat less in winter than in summer. These values, although not accurate, have some significance because two entirely different methods have yielded nearly identical results. They also substantiate the conclusion that the yearly

recession of the glacier front is slightly greater than the distance the ice in that part of the glacier moves each year.

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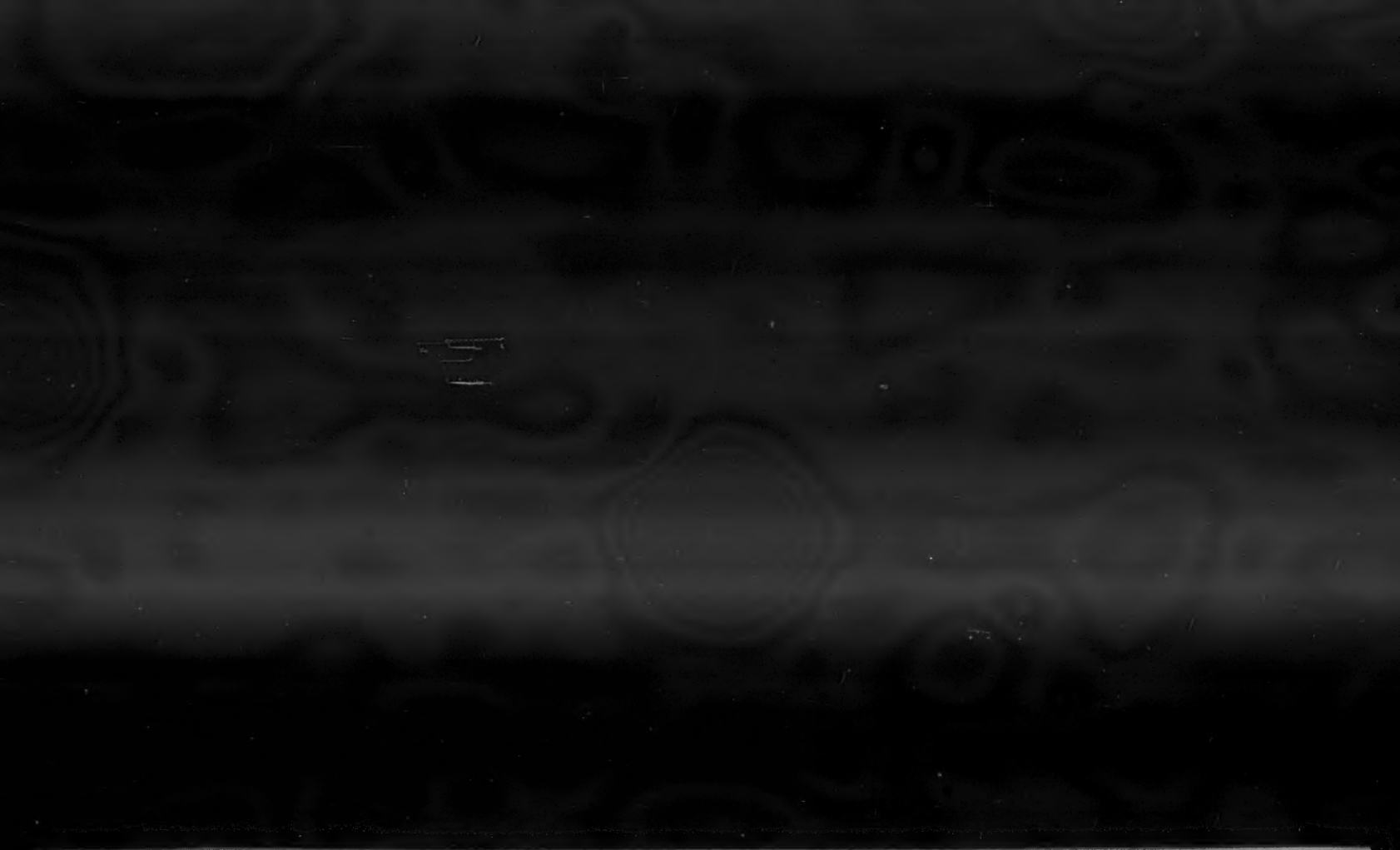

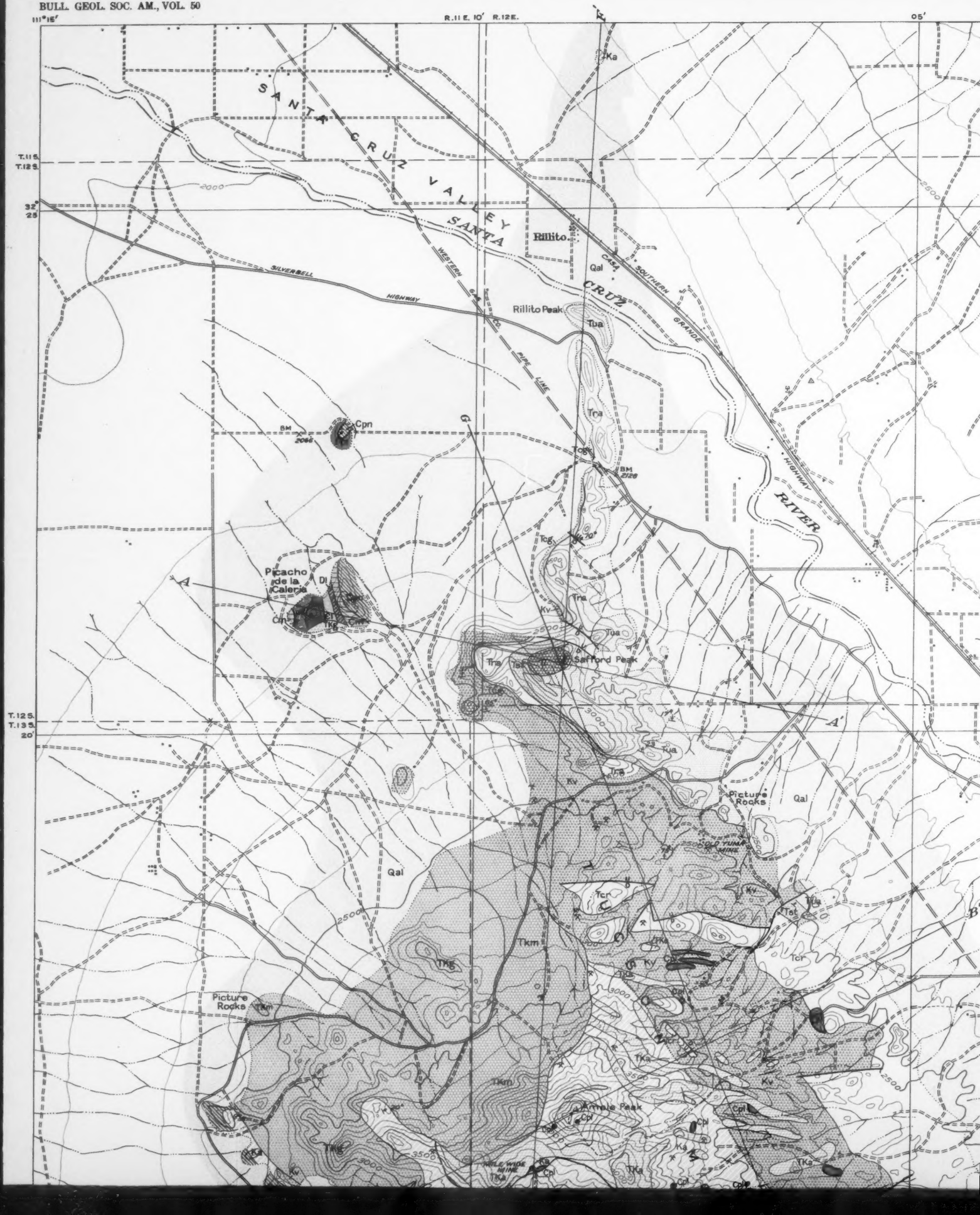
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25'
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WESTERN
SOUTHERN PACIFIC
GRANDE
SILVERBELL
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PIPE LINE
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Rillito Peak
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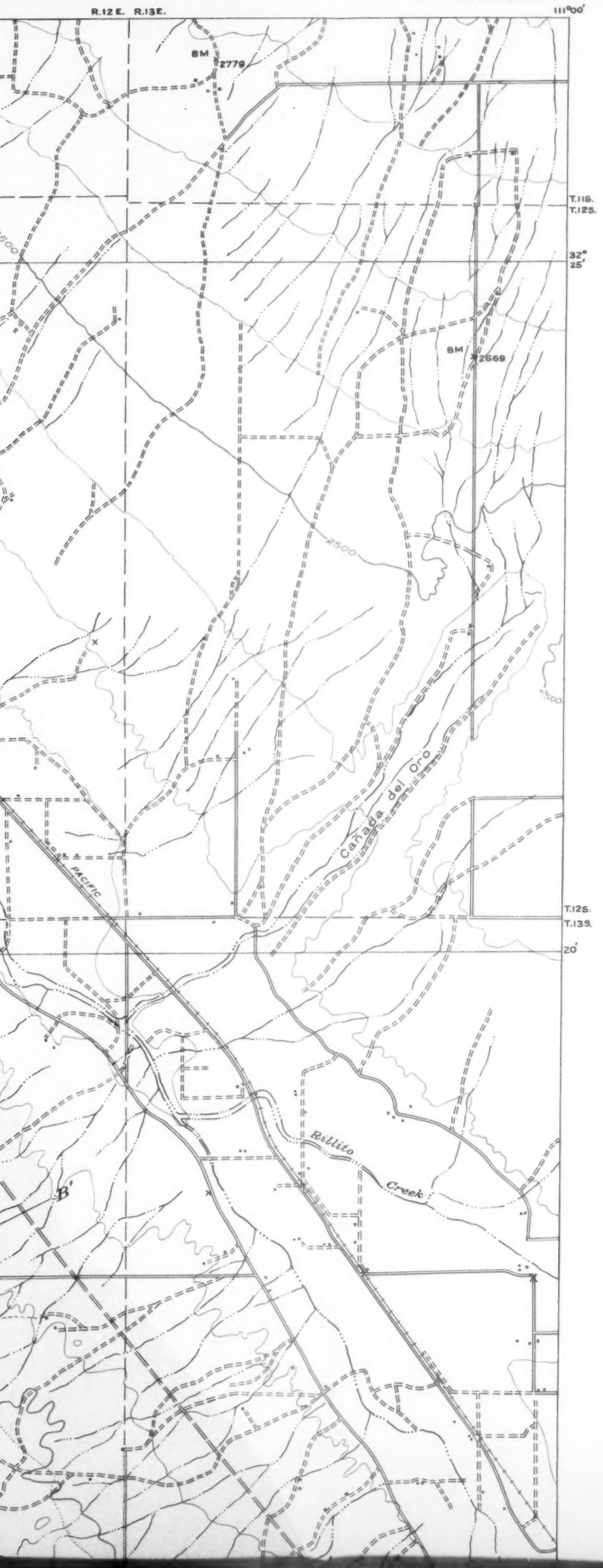


LEGEND

SEDIMENTARY AND EFFUSIVE ROCKS

QUATERNARY

Qal
Alluvium and talus

TERTIARY — CENOZOIC

TQ?b
Basalt
(Varied series of flows and one thin rhyolite tuff)

T?l
Lake beds
(Interbedded shales, limestones, and tuff)

SOUTHERN SECTION	CENTRAL SECTION	NORTHERN SECTION
Tsa **Shorts Ranch andesite** *(Massive andesite flows)*	Tua **Upper andesite** *(Massive andesite flows)*	Tua **Upper andesite** *(Massive andesite flows)*
Tbr **Biotite rhyolite** *(Massive rhyolite flow breccia)*		
Tda **Diopside andesite** *(Massive thin flow)*		
Tst **Safford tuff** *(Thin bed of glass shards)*	Tst **Safford tuff** *(Stratified water-lain tuff)*	Tst **Safford tuff** *(Coarse volcanic breccia)*
Tcr **Cat Mt. rhyolite** *(Rhyolite flows and mud flows)*	Tcr **Cat Mt. rhyolite** *(Rhyolite flows and mud flows)*	Tra **Rillito andesite** *(Massive andesite flows)*
		Tcg **Conglomerate** *(Thin reddish-brown conglomerate)*

CRETACEOUS — MESOZOIC

Ka
Amole arkose
(Arkose interbedded with shales and thin-bedded limestones)

Kr
Recreation red beds
(Massive, fine-grained, brick-red sandstone)

Kv
Cretaceous volcanic rocks
(Andesitic flows and tuffs)

CARBONIFEROUS

Cpl
Permian limestone
(Massive black limestone)

Cpn
Pennsylvanian limestone
(Massive gray limestone)

Cm

Picture Rocks
Qal
Tkm
TKs
Tcr
Kv
Ka
Cpl
TKa
Ti
Ts
Tsk
Tki
Ridge Mine
Ranger Station
Redmond Hills
Golden Gate Mtn
Snyders Hill
Saginaw Hill
ALTAPLANO
A/O ROAD
T.13 S.
T.14 S.
T.15 S.
B
C
D
E

Ka

Amole arkose

(Arkose interbedded with shales and thin-bedded limestones)

Kr

Recreation red beds

(Massive, fine-grained, brick-red sandstone)

Kv

Cretaceous volcanic rocks

(Andesitic flows and tuffs)

Cpl

Permian limestone

(Massive black limestone)

Cpn

Pennsylvanian limestone

(Massive gray limestone)

Cm

Mississippian limestone

(Massive gray limestone)

Dl

Devonian limestone

(Varied limestones, sandstone at base)

Cl

Cambrian limestone

(Thin-bedded limestone and sandstone)

Cq

Cambrian quartzite

(Massive vitreous quartzite)

pC

Pinal schist

(Fissile sericite schist)

INTRUSIVE ROCKS

Ti

Intrusives

(Acid dikes and volcanic neck)

Ts

Spherulitic rhyolite

(Sill-like masses of spherulitic rhyolite)

Tka

Amole latite

(Sills and irregular intrusive masses)

Tkm

Quartz monzonite

(Gray medium-grained east border of stock)

Tkg

Granite

(Coarse pink granite and porphyry west side of stock)

Tki

Minor intrusives

(Acid to intermediate small stocks and dikes)

U
D

CRETACEOUS — MESOZOIC

CARBONIFEROUS

DEVONIAN — PALEOZOIC

CAMBRIAN

PRE-CAMB. — PRE-PAL.

TERTIARY — CENOZOIC

Laramide ?

LATE CRETACEOUS – EARLY TERTIARY

MESOZOIC – CENOZOIC

10'

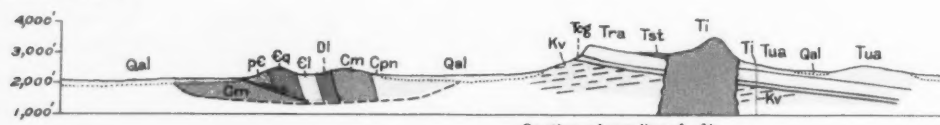
T.14 S.
T.15 S.

32°
05'

111°15'

R.11 E. 10' R.12E.

Topography by U.S.G.S.-C.W.A.
A. G. Kaetz, Supervisor
and W. Horatio Brown



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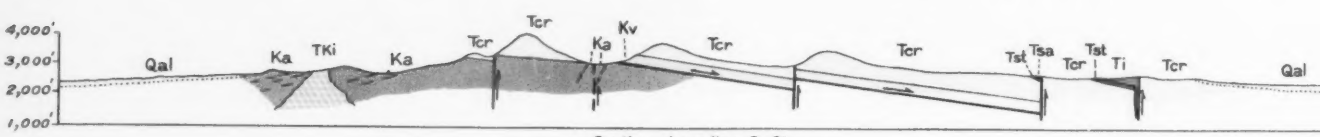

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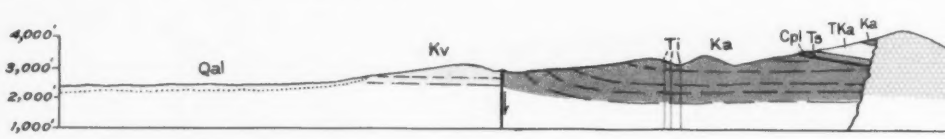

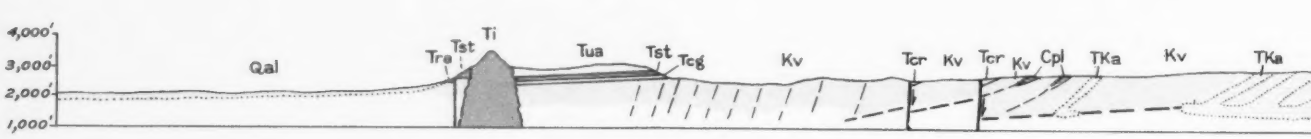

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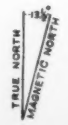
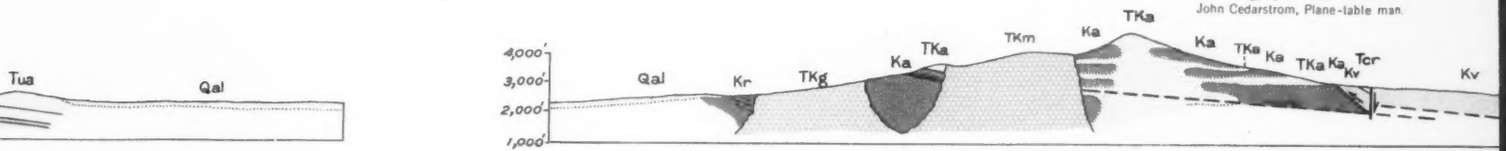


Geology by W. Horatio Brown
John Cedarstrom, Plane-table man.



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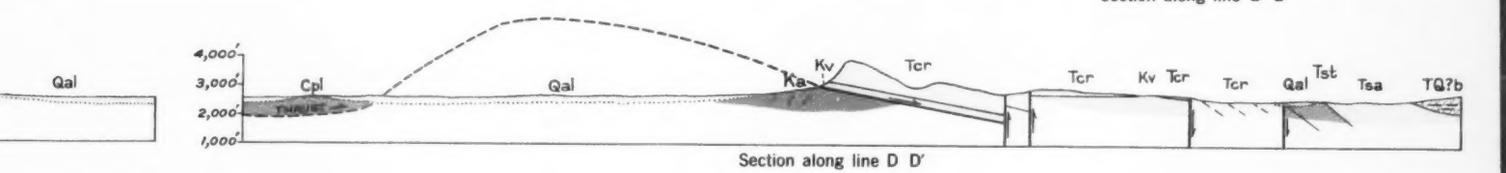

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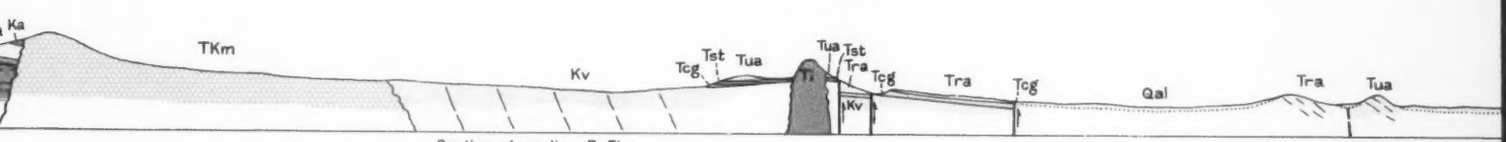

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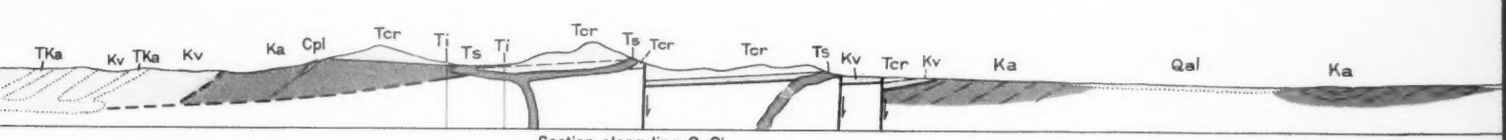

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C MAP AND STRUCTURE SECTIONS OF THE TUCSON MOUNTAINS, ARIZONA

Scale 1/62,500

1 0 1 2 3 4 Miles

5,000 0 15,000 Feet

Contour interval 100 feet

(Datum is mean sea level)

1939

OLOGIC MAP AND STRUCTURE SECTIONS OF THE TUCSON MOUNT

Scale 1/62,500

1 0 1 2 3 4 Miles

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Contour interval 100 feet

(Datum is mean sea level)

1939

Geology by W. Horatio Brown
John Cedarstrom, Plane-table map.

Tkm
Quartz monzonite
(Gray medium-grained east border of stock)

Tkg
Granite
(Coarse pink granite and porphyry west side of stock)

Tki
Minor intrusives
(Acid to intermediate small stocks and dikes)

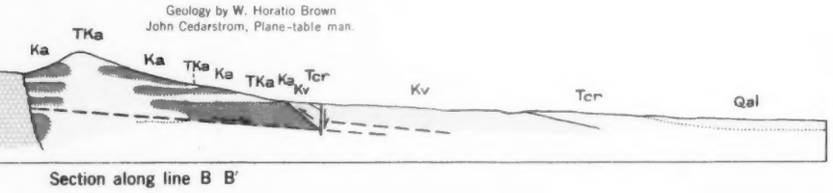
Laramide ?
LATE CRETACEOUS – EARLY TERTIARY
MESOZOIC – CENOZOIC

U
D
Fault
U, *upthrow* D, *downthrow*

Thrust fault
T *overburden side*
→ *dip of fault plane and direction of movement*

Strike and dip

Mine



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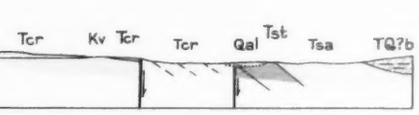

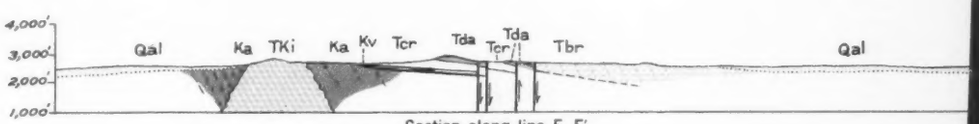

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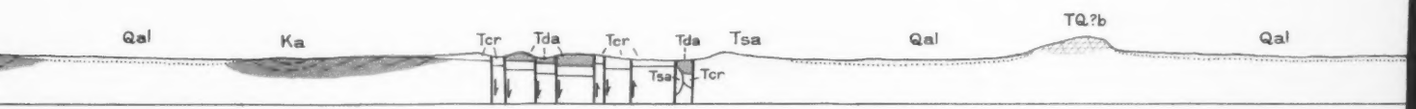

OUNTAINS, ARIZONA

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 50, PP. 697-760, 5 PLS., 2 FIGS. MAY 1, 1939

TUCSON MOUNTAINS, AN ARIZONA BASIN RANGE TYPE

BY W. HORATIO BROWN

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ABSTRACT

Tucson Mountains are an excellent example of the tilted, lava-capped ranges of southern Arizona. They consist of rocks comprising what appears to be one of the most complete columnar sections in the Basin Range province of Arizona. The study of the mountains has shown the presence of sediments ranging in age from Cambrian to Cretaceous, and Cretaceous, Tertiary, and Pliocene or Pleistocene effusives.

Laramide intrusives cut Laramide folds and a Laramide thrust fault, which is remarkable in its preservation. The fault was practically parallel with the early Tertiary erosion surface, and only a thin veneer of the overthrust block is preserved below the Tertiary lavas. The events in the historical sequence have been dated by the discovery of invertebrate remains in the Cretaceous sediments and plant remains in a Tertiary tuff.

The great reduction of the range by erosion and the burial of its outer edges have concealed the hypothetical Basin Range border faults, but the predominance of steep-angle normal faulting in the Tertiary rocks of the range indicates a similar structure of the buried borders. A complicated erosional history is indicated by the presence of pediments, probably of three different ages, and evidence for a former greater burial of the range in its own débris.

INTRODUCTION

PURPOSE OF THE INVESTIGATION

The study of the Tucson Mountains was started during 1925 when the writer was teaching at the University of Arizona. The range is adjacent to the University, and a knowledge of it was desired for the direction of student field work. It appeared to be one of the most typical Basin and Range mountains in the region and to have a complete columnar section.

The work was carried on for several days of each week for 4 years but was dropped when the writer left the university to engage in commercial work. In February 1934 a leave of absence was obtained, and with a grant from the Penrose Bequest of the Geological Society of America the work was resumed for 10 weeks.

In the following account of the geology an attempt has been made to summarize the knowledge of the area and especially to point out the problems that must be solved for a fuller understanding of its complicated history.

LOCATION AND EXTENT

The Tucson Mountains in southeastern Arizona occupy a large part of the eastern half of the Langhorne quadrangle, United States Geological Survey-Civil Works Authority topographic sheet, between latitude 32° 00′ and 32° 30′ North and longitude 111° 00′ and 111° 15′ West (Fig. 1).

The range trends about north-northwest and is bordered on the eastern side by the shallow Santa Cruz Valley and on the west by a similar valley locally called the Altar, Avra, or Abra valley. The city of Tucson is located at the east base of the central part of the range.

The range is 25 miles long and about 10 miles wide in the central part, narrowing toward both ends.

CLIMATE

The climate is semi-arid, with wide daily temperature range and low rainfall. Bryan (1925, p. 29) gives an extreme temperature range from

6 degrees to 112 degrees Fahrenheit and a mean annual temperature of 66.7 degrees at the Tucson station. The mean annual rainfall at Tucson is 11.85 inches but it is to a considerable extent concentrated in the July-August and December rains.

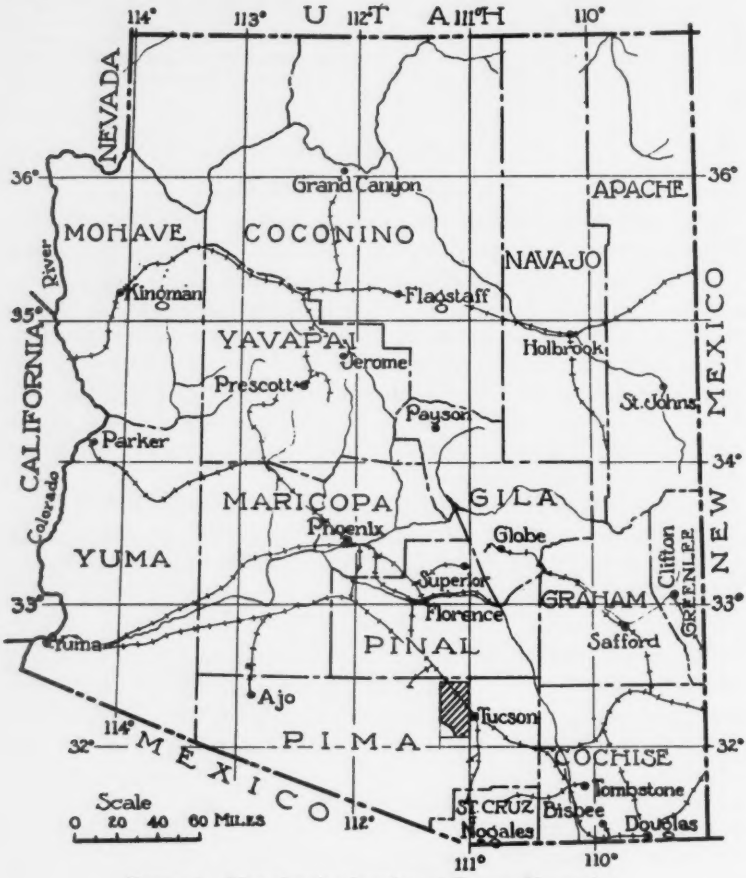

FIGURE 1.—*Map showing location of Tucson Mountains*

VEGETATION

Bryan, in his summary of the vegetation of the Papago country, has aptly described the region as an arboreal desert. The commonest trees and shrubs in the neighborhood of Tucson are the mesquite, palo verde, catsclaw, ocotillo, and palo fierro, which grow mainly on the pediments

and bajada slopes and are concentrated along the banks of the dry arroyos. The greasewood or creosote bush is the predominant shrub between arroyos. Perhaps the most striking plants are a large variety of the cacti, the most prominent of which is the giant cactus or sahuaro whose tall fluted columns will be noted in the illustrations. Smaller forms include the barrel cactus with its hook-shaped spines, the prickly pear, and the cholla. From the barrel cactus, water can be obtained. The cholla, especially the silvery cholla, which owes its color to its silvery spines, is the most annoying of these spiney plants. Members of the yucca family seem to be absent in the lower parts of the range but above about 3000 feet, especially on the slopes of Amole Peak, are fairly abundant, and several varieties occur.

Grass is scarce in the range and usually grows in small bunches several feet between centers. It is noticeably more abundant in the Desert Laboratory enclosure and in the higher, more inaccessible parts of the range. Whether this is due to less grazing or heavier rainfall is open to question. Certainly in the Desert Laboratory enclosure the abundance of grass, which is locally waist high along the arroyos, is due to protection from grazing.

ACKNOWLEDGMENTS

In the earlier stages of this work the writer had the advice of the late Drs. F. L. Ransome and W. M. Davis. To the latter especially is due much of the writer's interest in the problems of the Basin and Range structure and the problems of pedimentation. In the later stage of the work, Mr. John Cedarstrom operated the plane-table. Dr. A. A. Stoyanow kindly furnished manuscript sections of the Paleozoic sediments. Thanks are especially due to the writer's colleague, Mr. August Merz, Jr., for assistance in editing and to Professor B. S. Butler and Professor F. F. Grout for criticising the original manuscript.

The completion of the field and office work was made possible by a grant of $1000 from the Penrose Bequest of the Geological Society. Thanks are also due to Mr. Marvin Kleff, of Leadville, Colorado, for the use of his pantagraph in preparing the map.

PHYSIOGRAPHY

STRUCTURAL SUMMARY

As a background for a discussion of the topography it seems worth while to summarize the structures upon which it was developed.

The range consists of a series of folded and thrust-faulted Paleozoic and Mesozoic sedimentary rocks, upon the eroded and peneplaned edges of which lie a series of Tertiary effusives. A great thrust fault which can be traced almost the length of the range was nearly obliterated by

the erosion which left the overthrust block represented by a multitude of small klippen on the pre-volcanic surface. Post-volcanic block faulting, tilting of the composite mass toward the east, and erosion have given rise to an escarpment along the western side of the range and a dip slope on the east. Large intrusives in the west-central part, and transverse as well as longitudinal faulting throughout, have added to the complexity of the structure of the range.

PROCESSES OF EROSION

The processes of erosion prevalent in this semi-arid region have caused the retreat of the western escarpment of the Tucson Mountains, thereby developing a wide pediment and a main ridge that represents only a fraction of the former volume of the range.

Weathering in the range is almost entirely mechanical but varies with the kind of rock and with the altitude and relief. From bold cliffs, many-ton blocks are detached along joint planes and roll to the slopes or to the pediment below, breaking into fragments. This process is aided by sapping of the weaker rocks below or interbedded with the flows.

In some of the more massive formations exfoliation is taking place on subdued slopes. On very gentle slopes and on the pediment, rocks disintegrate chiefly by granulation.

The writer sees a distinct relationship between the color of the volcanic rocks and their susceptibility to granulation. In general the darker rocks seem to disintegrate much more readily than those lighter in color.

If this relationship exists—and it should be independently checked, as such an observation is subject to personal equation—it may well throw light on the causes of the granular disintegration. The darker volcanic rocks are usually the more basic, and the increase in volume due to a slight and almost imperceptible amount of oxidation of the iron compounds may be a cause of the breaking. The darker rocks absorb more heat, and temperature changes may be the cause of the breaking. If the writer's qualitative observations are verified, a further attack on the problem in the laboratory appears both justified and promising.

The only positive evidence of chemical weathering is the common solution-pitting and grooving of the surface of the limestone outcrops and a slight amount of oxidation of the ferromagnesian minerals. The prevalence of caliche may also be taken as evidence of chemical activity, as it hardly seems possible that all caliche was derived from limestone, especially that which occurs on slopes of volcanic hills remote from limestone outcrops.

The processes of weathering already discussed only loosen the material.

Once the fragments have been detached and rolled down the slopes, transportation by intermittent streams begins. Stream action predominates beyond the base of the escarpment, and the lower boulder-strewn slopes are trenched by steep cloudburst furrows with embanked levee-like sides. In these furrows large boulders are carried.

On the gently sloping pediment, however, the carrying power of the streams is diminished, and the larger material has to be comminuted before transportation. A line of large boulders is common at the head of the pediment below the escarpment.

The arroyos of the range usually head in the cloudburst furrows. At the foot of the escarpment they change to entrenched valleys without levees; as they extend out on the piedmont slopes the height of the banks decreases until they begin to branch indefinitely on the alluvium of the intermont plains.

Abrasive wind action appears to be unimportant in the area. Desert varnish is common, and it is inconceivable that desert varnish can form and persist in an area of abrasive wind action. Large quantities of dust, however, are undoubtedly removed by wind during the spring months, as several "dust devils" or whirlwind dust storms can frequently be seen on the pediments and on the plains simultaneously.

The weathering of limestone presents a somewhat different problem from that of other rocks. The widely recognized tendency of limestone to form hills in the desert can be seen throughout the range. This might be attributed to lack of rainfall and to the prevalence of mechanical weathering in the desert. However, observations described later make the writer doubt the validity of this explanation.

In the conglomerate in the Recreation red beds, limestone pebbles are common and are associated with pebbles of andesite, quartzite, and chert. On the Red Hills these limestone pebbles are the only appreciably weathered ones in the conglomerate. They are being removed by solution which first attacks the periphery of the pebbles and forms a trench in the limestone next to the matrix. At this stage the pebbles are still convex and marked by minute radiating channels extending down to the peripheral trench next to the matrix of the conglomerate. Study of a series of pebbles shows that they are gradually being reduced to final spire shapes whose channeled slopes are distinctly concave. Thus, solution attacks limestone more rapidly than mechanical weathering attacks the associated rocks. It seems that the differential rate of mechanical weathering and solution does not fully explain why limestone forms hills in the desert.

Two features of limestone hills may explain their resistance. Boulders are often found cemented onto limestone outcrops by caliche. This

typical desert deposit is common near the surface and extends to depths as great as 50 feet. It cements all the porous materials in which it occurs. Also, around the base of limestone hills loose debris or talus is usually missing. In its place is often a breccia so thoroughly cemented by caliche that it serves as a buttress against erosion. These two features will probably, at least partly, account for the hill-forming tendency that cannot be accounted for by differential weathering alone.

STAGE AND FORM

General statement.—Erosion has reduced the range to six distinct physiographic elements: the Altar alluvial embankment, the Altar pediment, the western escarpment, the dip slope, on which there is minor pedimentation, the basaltic mesas, and the Santa Cruz alluvial embankment.

Pediments.—The pediment on the western side of the range slopes westward from the base of the escarpment for, at most, 3 miles. This rock-floored plain is carved chiefly on Cretaceous sediments, and in areas of almost perfect outcrops single beds may be traced for considerable distances around the noses of successive folds.

Areas of almost perfect exposures are not present throughout the area shown on the areal map (Pl. 1) as rock floored. The writer has shown the extreme limits of outcrops and realizes that the percentage of pediment or exposed rock floor is greatly exaggerated. It will be noted that the writer's use of the term pediment is not in perfect accord with the definition by Bryan. The writer is inclined to limit the term to that part of the slope which is definitely rock floored, because of the usual impossibility of determining the depth of the alluvial cover. However, the exaggeration in mapping brings the map nearly in accord with the original definition. In general, rock floors seem to be confined to slight elevations on the piedmont slope and to embayments in the escarpment. Alluvium covers the intervening areas to unknown depths. The slight elevations seem to be extensions of spurs on the escarpment, and the intervening alluvial areas appear to be extensions of valleys in the escarpment. If this correlation is correct the alluvial areas may cover the bedrock to considerable depth and should not be classified as pediments.

The pediments vary greatly. Undissected rock floors at base level with the alluvium contrast with dissected and entrenched pediments.

At the southern end of the western side of the range, south of the Ajo road, stretches a well-developed rock-floored plain carved on steeply tilted and folded Cretaceous sediments. This surface is best developed along the base of the escarpment; from half a mile to a mile beyond it is usually concealed. Islands of bedrock usually the extensions of spurs rise slightly through the alluvium. They are almost plane surfaces but are definitely

convex and are apparently miniature examples of Davis' (1933, p. 211) domes or Lawson's (1915) panfans. They are so low and so near the base level of the surrounding alluvium that only sheetwash and occasional rills affect them. Here the most noticeable relief, 6 inches to 1 foot, is caused by thin-bedded limestone in the shales and arkoses.

Between the Ajo road and the Piedmontite Hills on the western side of the range a narrow dissected pediment extends along the foot of the escarpment. The depth of dissection becomes progressively greater toward the north. The alluvium that caps the flat-topped surface of the plain between arroyos to a depth of several feet was washed down from the escarpment and, toward the north, from the high country around Amole Peak. The arroyos are about 15 or 20 feet deep and are superimposed streams cut in truncated Cretaceous sediments.

West of the Piedmontite and Red Hills there are no pediments, and the alluvium of the Altar Valley laps up on the foot of the range, but to the north, in the areas of granitoid intrusives, extensive pediments have been formed. Some are of the normal type, similar to those farther south, and consist mainly of rock-floored plains skirting the base of the range. Others are plains formed when intermittent westward-flowing streams widened their lower courses. Erosion has cut a distinct plain in the granite between the limestone hills at the extreme western end of the range and the granite hills to the east. Furthermore, a pediment is encroaching at the mouth of the valley just north of the Red Hills.

On Cretaceous volcanic rocks between the granitoid intrusives and Safford Peak are extensively developed pediments, one of which slopes from the main mountain axis toward Altar Valley. A narrow, poorly developed pediment slopes south from the base of the escarpment south of Safford Peak. These pediments, if considered as separate units, merge opposite the pronounced gap west of the Old Yuma mine and form a distinct embayment.

Between Safford Peak and the Santa Cruz River, no pediment is exposed along the western side of the range. If one is present it is buried to an unknown depth. Here coarse and fine alluvium merge on the floor of the plain at varying distances from the foot of the escarpment. It is a relatively undissected plain sloping gently toward the Altar River and is broken only by two prominent limestone hills.

Rock-floored plains are to be seen at only a few points on the eastern side of the range. Between Safford Peak and Picture Rocks a small plain has been carved on the tilted Upper Andesite. In the embayment south of Tumamoc Hill is a pediment, dissected apparently before complete development. Several streams with relatively wide flat valleys drain

the dip slope between these two points. The wide valley bottoms possibly represent an earlier incompleted stage of pedimentation.

Farther south, cutting the Tertiary volcanic rocks nearly at the base level of the overlapping alluvium, pediments are extensively developed similar to those on the western side.

At both ends of the Black Hills pediments are poorly developed. Judging from several level outcrops of basalt near its entrance, San Xavier Mission apparently stands on one. The small basaltic hills between the mission and the Black Hills rise above the Santa Cruz alluvium as residuals on this rock floor. At the southwest end of the Black Hills a pediment only a few hundred feet wide is exposed between the hills and the large wash carrying granitic detritus. Presumably the granitic alluvium buries this pediment.

In general, pediments are better developed at the foot of the escarpment on the western side of the range than they are on the dip slope. Whether this is due to structure, difference in bedrock, or difference in the erosion and maintenance of alluvial base level by the two rivers is open to question. Until the physiographic history and the relationship of pediments and alluvium are more completely deciphered, it is unwise to attempt conclusions.

In the Tucson range, two features of pediments seem especially important. First, all have probably been deeply buried by alluvium whose present thickness is no measure of the former depth of burial. The removal of the alluvium from such a relatively flat surface as a pediment would proceed at a steadily decreasing rate.

Second, there is a definite relationship between the angle of slope of pediments and the present extent of burial. The dissected pediments with a thin veneer of gravel have slopes averaging about 200 feet to the mile. The undissected pediments, such as those south of the Ajo road, which are extensively covered with alluvium and on which the rock-floored surfaces are planed nearly to an alluvial base level, have flat grades of about 50 feet to the mile. Alluvial plains, which may cover problematic buried pediments such as that in the extreme northwest part of the area, have still gentler gradients.

The writer believes that he can see a similar relationship in the pediments described by others. Two pediments are illustrated by Bryan (1925). The one north of the Sacaton Mountains has a gradient of about 50 feet to the mile, and "the northern part of the area has a heavy burden of alluvium." The pediment on the northern border of the Baboquivari is dissected and has a steeper slope—about 100 feet to the mile—and the one at Fresnal 80 to 100 feet per mile in the dissected portion. The gravel-

veneered dissected pediments in Utah, described by Rich (1935, p. 1014), have gradients of about 200 feet to the mile.

There has been much discussion of the origin of pediments, but no general agreement. Until more data are available on the shape of the rock floors and the depth of the alluvial cover [1] in many typical pediments, the writer is inclined to reserve judgment on origin. For the time being, the broad generalization of Bryan (1925, p. 96) that pediments are formed by

> "(1) lateral planation by streams issuing from the canyons, (2) rill cutting at the foot of the mountain slopes, (3) weathering of outliers and unreduced remnants, with transportation of the debris by rills,"

seems the best working hypothesis.

Escarpment.—Where controlled by the eastward-dipping lavas at both its northern and southern ends the western side of the range is marked by a prominent escarpment which is very straight and uniform. Areas of intruded and metamorphosed Cretaceous rocks cause spurs in the escarpment, for example Golden Gate Mountain, and indicate the former greater extent of the range.

In the west-central part of the range, however, the escarpment gives way to a single mountain mass climaxing in Amole Peak. This is the highest point on the range and is composed of granitoid and porphyritic rocks intruded into sediments. Drainage from this mass is largely radial, although the line of the escarpment is separated from the peak by a high small saddle on the south and a very large low saddle on the north.

At the saddle 2 miles southeast of the Peak the escarpment-forming lavas dip away from the peak (Pl. 2, fig. 2). Remnants of them dipping away from the peak occur on the divide 1¾ miles north of the peak. The lavas extend continuously around the east side of the peak. It seems almost certain, therefore, that lavas covered Amole Peak and that the range was a completely lava-capped block, with a continuous escarpment which lay miles west of its present position.

The escarpment north of Amole Peak faces southwestward, but near Safford Peak it reaches its highest elevation and turns abruptly northward. East of the turn and forming part of the escarpment and part of the dip slope area is the prominent spire of Safford Peak, a volcanic neck.

Gaps or passes through the escarpment are visible for miles. This serrated skyline has led to the nickname, "The Cardboard Mountains"; for in profile at sunset the peaks stand out as though cut from cardboard. Most gaps are the site of cross faults, some of which displace the escarpment although they are difficult to locate because they are covered by

[1] It seems probable that such data can best be obtained by geophysical methods.

alluvium. Pediments usually encroach strongly on these zones of weakness. Superficially the gaps resemble wind gaps, but they have almost certainly been formed by streams working headward from opposite sides of the range.

Dip slope.—The eastern dip slope drains to the Santa Cruz River through intermittent streams which are deeply entrenched in the lavas. The streams were probably originally consequent on the tilted block. Many of them follow fault zones and trend south of east down the dip slope toward the north to northwest-flowing Santa Cruz River. It seems remarkable that examples of stream piracy are not more common and that this anomolous drainage trend has persisted in spite of the great reduction the range has undergone since tilting.

The most widespread formation on the dip slope is the Cat Mountain rhyolite, but it seems certain that the Shorts Ranch andesite once covered practically all of the region from the Ajo road to a point north of Twin Hills. Its down-faulted inliers are numerous and its thickness in this part of the range is difficult to determine, although it is at least 500 to 1000 feet and probably more. It is impossible to say whether any younger volcanic rock overlay the andesite in the central part of the dip slope.

North of Twin Hills, along the contact between the Cat Mountain rhyolite and the alluvial embankment, the alluvium is composed almost entirely of Shorts Ranch andesite in large angular fragments so numerous that they resemble andesite in place. The source of the blocks is unknown. If they came from Twin Hills the condition of drainage and burial by alluvium was very different from the present. If they are of local origin the andesite formation has recently been removed from above the rhyolite. The Safford tuff seems to be missing from the debris, and, in view of the example of deep burial cited in succeeding paragraphs, the writer is inclined to believe that the andesitic material was derived from Twin Hills during a former erosion period.

The control of the drainage by the easterly trending faults unbalances the main Santa Cruz drainage. Normally the tributaries would flow north or northeast toward the Santa Cruz River. This condition has favored stream piracy, in which streams cutting headward through the rhyolite ridge along the northeast border of the range have captured streams flowing along the cross faults.

For about 7 miles along the northeastern side of the range, erosion has cut through the capping of Tertiary lavas and exposed the Cretaceous rocks (Pl. 1, sec. B-B′). This has produced on the dip slope an escarpment with a distinct depression between it and the main range. The cause of this depression is not clear, but the fact that the Twin Hills fault is

FIGURE 1. TILTED BLOCK STRUCTURE
Golden Gate Mountain in center, Amole Peak left background.

FIGURE 2. WESTERN ESCARPMENT
Shows component of dip of lavas away from Amole Peak.

WESTERN SIDE OF RANGE

FIGURE 1. COARSE SAFFORD TUFF
From vicinity of Safford neck.

FIGURE 2. FINE-GRAINED WATER-LAIN SAFFORD TUFF
Near Twin Hills.

in line with it is suggestive. Shearing on the strike of the Twin Hills fault was observed in the Cretaceous volcanic complex, but whether the fault extends through the Cretaceous belt was not determined. Presumably, any fault that extends along this line would have to be offset on the fault that bounds the Cretaceous area on the southeast.

Other streams on the dip slope have apparently been superimposed on the lavas after a much more complete burial of the flanks of the range by the alluvium than exists at present. Three and a half miles northeast of Amole Peak, at the edge of the area of bedrock, there is a prominent little hill composed of the Upper Andesite. Along the northwest side of the hill is an arroyo, about 25 feet deep, cut into the andesite. Northwest of the arroyo the andesite does not extend to the surface but is covered by alluvium which mantles the flat-topped ridge. The alluvium is not composed of local rocks, but of rocks from the slopes of Amole Peak—Amole Latite and metamorphosed Cretaceous sediments. Apparently the lavas of this area were once almost entirely buried by alluvial fans which extended well up on the slopes of Amole Peak. This relation and that at Twin Hills seem to point to an earlier period of intense alluvial deposition, during which the range was almost completely covered by alluvium.

The influence of faults on the drainage pattern is brought out in the basin north of Cat Mountain. There, within half a mile, three prominent faults, and probably twice that number of unmapped minor ones, intersect. They have so weakened the rock that a large basin has been eroded at the head of San Juan Wash, which carries the drainage of the area eastward to the Santa Cruz River. This wash follows a large and persistent fault and throughout most of its course is deeply entrenched. The basin at its head occupies an area of about a square mile and a half. It is maturely dissected and rolling. It contrasts sharply with the stream valleys on the dip slope that have been widened by the encroachment of the pediments. These are, in general, U-shaped and somewhat wider at the mouth than at the head. In many places noted in the range, pedimentation in these valleys on the dip slope has ceased, and the streams in them have been entrenched, initiating a new cycle of pedimentation. Perfect pediments occur in the southern part of the range, on the dip slope near Beehive Peak, which are at or near the base level of the alluvial embankment.

FORMATIONAL UNITS

COLUMNAR SECTION

Nearly every period represented in the rocks of southern Arizona is included in the columnar section of the range (Fig. 2). In Figure 2 the thickness of the formations is shown to scale, and the unconformities are

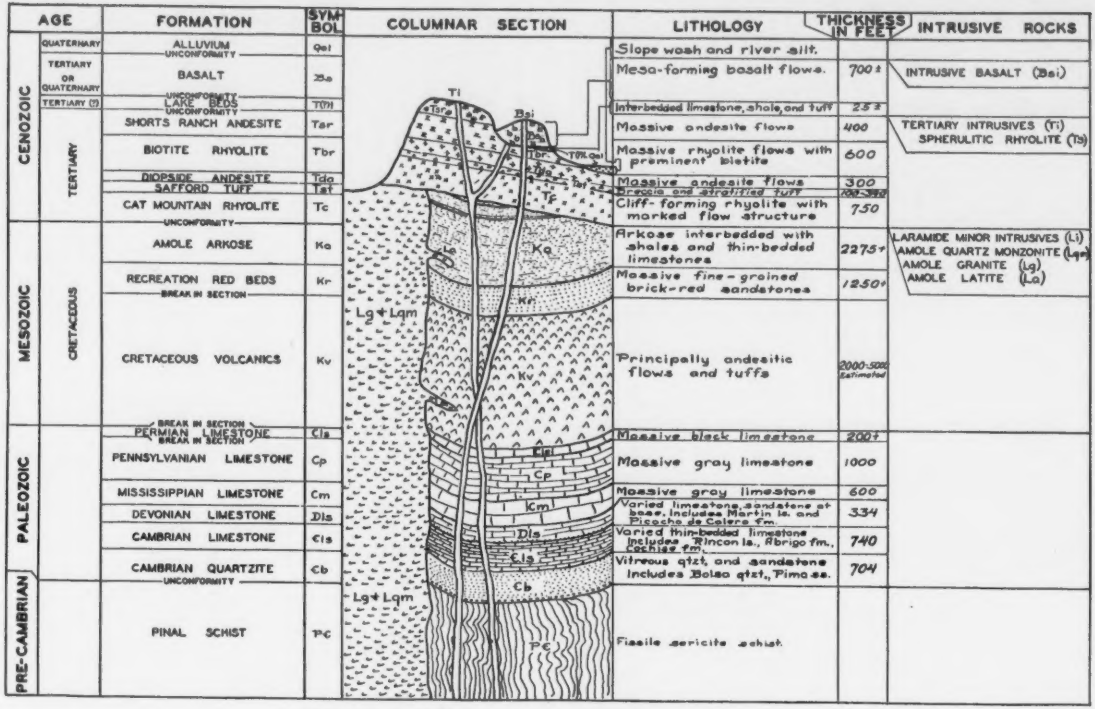

FIGURE 2.—*Columnar section of Tucson Mountains*

diagrammatically portrayed. Because of difficulties in representation, only the sequence of lavas in the southern part of the range is included. The intrusives are shown diagrammatically, to give the age relations.

PINAL SCHIST

Except for small blocks scattered along the great thrust, Pinal schist outcrops only on the western side of Picacho de Calera Hills. It unconformably underlies the Bolsa quartzite and is largely covered by slide rock and talus from that formation. Because of its position beneath the Bolsa quartzite and its lithologic similarity to the Pinal schist of central Arizona it is tentatively correlated with that pre-Cambrian formation.

The schist is very fine grained and has a satin-like sheen; quartz and sericite can be recognized with the hand lens. The weathered schist is light gray-green. No fresh material was found. The cleavage is nearly vertical. The rock is friable and weathers to a powder.

PALEOZOIC ROCKS

Except for scattered small blocks, practically all the Paleozoic formations in the range are confined to a section in the Picacho de Calera Hills, west of the northern end of the range. The western hill is composed of Pinal schist and Bolsa quartzite. The eastern hill is underlain by Mississippian and Pennsylvanian limestones. Cambrian and Devonian limestones occur in the intervening saddle. The section from Pinal schist to Pennsylvanian limestone is continuous and relatively undisturbed.

This locality has been studied and the section measured by Stoyanow (1936) and his students, and he has generously furnished the Paleozoic sections for this paper.

CAMBRIAN SYSTEM

Cambrian formations are known only in the Picacho de Calera Hills and possibly in a few small blocks of thin-bedded limestone that resemble the Cambrian in the complicated thrust-fault zone. Lithologically the Cambrian rocks can be divided into two principle members, a lower quartzite and an upper limestone. On a paleontologic and lithologic basis, Stoyanow has divided them into five members, as follows:

	Feet
Rincon limestone. Cliff-forming, pink, mostly crystalline limestone. Abundant trilobites	40
Abrigo formation. Thin-bedded, cherty dolomitic limestone. Upper Cambrian trilobites	400
Cochise formation. Varicolored sandstone and shale. Thin-bedded, micaceous, gray sandstone, blue algal limestone. Middle Cambrian trilobites	300
Pima sandstone. Buff, hard sandstone with Middle Cambrian brachiopods	4
Bolsa quartzite. Basal conglomerate, pebbly grits, cross-bedded, vitreous quartzites	700

These rocks form the western hill and the saddle between the two hills. They are cut by two dikes or sills.

DEVONIAN SYSTEM

Devonian outcrops are confined to the eastern side of the saddle between the two Picacho de Calera Hills. The following section described by Stoyanow is well exposed.

	Feet
MARTIN LIMESTONE	
1. Massive, hard, gray limestone; fossils scarce	150
2. Yellow and brown calcareous sandstone	25
3. Gray limestone weathering to dark brown, with bands of chert at base and containing abundant brachiopods	80
4. Gray limestone with traces of brachiopods	6
PICACHO DE CALERA FORMATION	
1. Brown calcareous sandstone replete with fish teeth	2
2. Black dolomite	25
3. Yellow crystalline limestone largely made of small calcified algal bodies and interbedded with thin, flaggy, blue limestone; small goniatites are sporadically found	2
4. Blue limestone in beds 2-4 feet thick, composed of large spherical stromatoporoids and algae with abundant, but poorly preserved, zaphrentoid and favositoid corals	40
5. Yellow calcareous sandstone with well-rounded sand grains, probably of subeolian origin	4

MISSISSIPPIAN SYSTEM

Mississippian rocks also seem to be confined to the Picacho de Calera area, although because of lithologic similarity they may have been missed in the area of scattered limestone blocks along the great thrust-fault zone. At the Picacho locality there are two separate areas of these rocks. No detailed section of the Mississippian is available, but Stoyanow (personal communication) gives the thickness as 600 feet and correlates it with the Escabrosa. The principal and complete section forms the western slope of the eastern hill. A smaller area occurs at the western end of the hills in fault contact with the Pinal schist.

The Mississippian is a light- to dark-gray limestone containing shaly beds. The more rapid weathering of the shaly beds gives the outcrop a ribbed appearance.

PENNSYLVANIAN SYSTEM

The principal occurrence of the Pennsylvanian is also in the Picacho section on the eastern end of the eastern hill. It is a limestone very similar to the Mississippian but not so dark gray and contains some pink beds and chert. It is more granular than the Mississippian. Crinoidal beds occur in the section. No detailed section is available, but Stoyanow (personal communication), who correlates it with the Naco, gives the thickness as 1000 feet.

The Pennsylvanian also occurs in the small limestone hill 2 miles north of the Picacho, where it is thrust onto the Permian. In the hills that form the westernmost extremity of the range, Pennsylvanian limestone is in contact with Cretaceous volcanic rocks, but the whole area is so highly

metamorphosed by the nearby granite that the relations are obscure. Blocks of Pennsylvanian limestone are probably scattered along the great thrust fault, but the writer has not been able to find any.

Associated with the limestone blocks in the great thrust zone are several small unmapped quartzite blocks. This quartzite is almost white, well sorted, and very clean. It is quite different from the Bolsa. It probably represents a Pennsylvanian horizon higher than that at the Picacho de Calera Hills, although no fossil evidence has been found. Stoyanow has found similar quartzites higher in the Pennsylvanian elsewhere in the region.

PERMIAN SYSTEM

The Permian rocks of the range have recently been assigned to the Snyder's Hill formation by Stoyanow. Snyder's Hill is on the Ajo road, 3 miles west of Cat Mountain. It consists of dark limestone folded into a syncline. The thickness exposed is estimated at about 200 feet, but neither the top nor the bottom of the formation is exposed.

Among the limestone blocks on the great thrust those containing the typical fauna of the Snyder's Hill formation are most abundant. Especially fossiliferous blocks are found along the base of the escarpment near the Silver Lily dike system.

CRETACEOUS SYSTEM

General statement.—The most abundant sedimentary rocks in the range are of Cretaceous age. Three units are easily recognized. The upper, or Amole arkose, formation is most prominent and forms the large syncline on the western pediment. The Recreation red beds underlie the Amole arkose and occur chiefly on the western side of the syncline in the Red Hills. The lower volcanic member of the series occurs chiefly in the north-central part of the range in fault contact with the other members.

Volcanics.—Cretaceous volcanic rocks are widespread. The largest area extends along the western side of the range, from the granitoid intrusives nearly to Silver Bell pass, around the northern side of Amole Peak and along the eastern slope. Another belt extends along the base of the western escarpment, from the vicinity of Amole Peak nearly to the southern end of the range. In this belt, much Amole arkose is mapped with the volcanics. A stratigraphically important area is on Piedmontite Hill. Other outcrops are scattered along the southeastern side of the range.

Except in the area at Piedmontite Hill the Cretaceous volcanic series is so complex structurally that no attempt has been made to measure its thickness. An estimate of 2000 to 5000 feet recognizes the steep dips in the tuffs and the large areas covered.

Nearly all members of the formation have one common characteristic; they are dark purplish gray to purplish black on weathered surfaces except

where unusually altered. This contrasts sharply with the various shades of brown of the Tertiary volcanics.

Included in the series are flows, predominantly basic andesites but also dacites and latites, rhyolitic tuffs, partly water-laid tuffs, and breccias of andesitic material. Except for the rhyolite tuff, all members have two distinguishing features. The groundmass of the flows consists of a felt of small plagioclase crystals, and the plagioclase phenocrysts usually range from andesite to labradorite. The ferromagnesian minerals are usually altered beyond identification.

In most places four members constitute most of the formation. The commonest is porphyritic andesite with prominent phenocrysts of plagioclase half a centimeter in length. Weathered surfaces are purplish black, and fresh surfaces are a trifle lighter. In thin section the plagioclase phenocrysts show autoclastic brecciation. The ferromagnesian minerals are completely altered but have outlines suggesting pyroxenes. The relations of this andesite to the other members of the group are not clear. Usually it appears to be a flow, and exposures north of the Old Yuma mine are vesicular; locally it may be intrusive.

The andesite is associated with purple quartz latite porphyry containing short plagioclase and orthoclase and inconspicuous quartz phenocrysts. It too weathers to purplish black. Fragments of this rock are common in the volcanic breccias.

Coarse volcanic breccias consisting chiefly of fragments similar to the flows crop out over wide areas. Most of them appear to be flow breccias. They have the usual blackish-purple colors. Interbedded with the flows and breccias are well stratified and sorted arkoses or tuffs, clearly of water-laid origin. They consist of fragments of plagioclase and of the characteristic groundmass material of the volcanic rocks, with minutely matted plagioclase crystals. No shards of pumice have been observed in these beds. All the dips shown in the areas of these volcanics were taken on these beds. The partial section of the volcanic series is given (Table 1), not because it is typical but because it seems to show a relationship between the volcanic rocks and the Recreation red beds.

TABLE 1.—*Piedmontite Hill section of part of Cretaceous volcanic series*

Top of hill and section	Feet
1. Light pinkish-white dense rhyolite. Occasional inclusions altered to piedmontite. Under the microscope the rock is seen to consist of devitrified shards	50
2. Brownish-purple andesite tuff with arkosic groundmass, pebbles 2 to 3 inches in diameter	15
3. One- to three-foot reddish-brown arkose, well stratified and locally cross-bedded, underlain by andesitic tuff with pebbles 2 to 3 inches in diameter	29
4. Similar arkose underlain by similar tuff ranging from brown to dark purples characteristic of the series farther north	46

Top of hill and section

	Feet
5. Similar red arkose bed underlain by tuff. All arkose beds cap small cliffs. Locally generally coarser arkose grades toward that in the Recreation red bed series. In the tuffs the andesitic pebbles vary from 1 to 8 inches in diameter and are locally larger. The pebbles range from rounded angular to well rounded and from scattered to closely packed distribution in an arkosic matrix	31
6. Purplish-gray andesitic flow with scattered patches of piedmontite	15
7. Fine-grained red sandstone apparently identical with the Recreation red beds	26
8. Zone of purplish-gray arkosic tuff and interbedded andesites, with considerable alluvial cover	45
Fault	

The most significant feature of this section is the occurrence of red beds of the Recreation red beds type, interbedded with the volcanic rocks. The interbedding of conglomerate, consisting chiefly of typical andesitic material from this series, with the red beds suggests that the volcanic rocks are below the red beds in the section and that the change is gradational.

The metamorphism of the volcanic rocks is of three types: Pervasive high-temperature metamorphism which has altered the ferromagnesian minerals to epidote, piedmontite, vesuvianite, and other silicates; hydrothermal alteration, especially in the area west of the Old Yuma mine; and baking at the contacts with the overlying Tertiary volcanic rocks.

Intense jointing during Laramide and Tertiary diastrophism weakened the flows, and they usually form either well-developed pediments or zones of sapping along the bases of escarpments.

The evidence for the age of the volcanic rocks is discussed on a later page. Undoubtedly they belong with the pre-Laramide series as they unconformably underlie the Tertiary volcanic rocks near Safford Peak. Superficially they resemble the volcanic rocks near Aravaipa which were classified with the Cretaceous by C. P. Ross.

Recreation red beds.—Underlying the Amole arkose is a formation called the Recreation red beds because of its color and its occurrence in the Tucson recreational area. It is practically confined to the Red Hills and numerous, small, poorly exposed outcrops in the zone of the great thrust. These latter outcrops are common in the area of the thrust east of Amole Peak but are too small to be shown on the map.

The formation consists of a series of uniformly brick-red, fine-grained sandstones. Under the microscope they are seen to be even grained and colored by a ferruginous cement. Some variation can be recognized, and the uniformity is broken by a conglomerate consisting of pebbles of Cretaceous andesite. The conglomerate ranges from 5 to 30 feet in thickness. The section measured in Gould Gulch, at the south base of the Red Hills, is nearly on the strike of the fault that follows the east base of the Piedmontite Hills. Although the fault was not recognized in the section, a

peculiar type of jointing found near the middle of the formation may reflect its presence.

TABLE 2.—*Gould Gulch section of Recreation Red Beds*

	Feet
AMOLE ARKOSE	
Transition zone. Massive gray to purplish-gray arkoses, well bedded. A few 5- to 10-foot beds of brick-red fine-grained arkose. Base taken at base of gray arkoses.	
RECREATION RED BEDS	
1. Brick-red to dark brick-red fine-grained arkose. Some zones of limestone pebbles or concretions and a 1-foot limey bed near base. Locally reduced to pink along yellow carbonate veins	200
2. Purplish-gray conglomerate composed of pebbles half an inch to 6 inches in diameter, angular to well rounded. Pebbles chiefly Cretaceous andesites, together with limestone, quartzite, and chert. The matrix is chiefly purplish arkose, and some 6-inch to 1-foot arkose beds occur in the conglomerate	15
3. Monotonously massive, fine-grained, brick-red arkose	650
4. Coarse gray to reddish purple arkosic and locally tuffaceous-looking beds up to 10 feet thick, interbedded with fine-grained brick-red beds of the type in Bed 3. Locally cross-bedded. Coarse fragments in one bed up to half a centimeter in diameter, definitely andesitic	400
Base covered by Quaternary alluvium	

Although the red beds are limited areally, they form hills, whose lower slopes are eroded into typical entrenched pediments.

The red beds are overlain conformably by the Altar arkoses. The base of the series is covered by valley fill and, although nowhere found in contact with any older formation, the Cretaceous volcanic series probably lies below. For example, the conglomerate in the red beds is composed predominantly of andesitic pebbles and cobbles of the same types as the Cretaceous volcanic rocks and is practically indistinguishable from the tuffs or breccias in the Cretaceous rocks of Piedmontite Hill and the large volcanic area north of Amole Peak. Furthermore, the Cretaceous volcanic series of Piedmontite Hill contains beds of red arkose which are indistinguishable from the Recreation red beds.

Because of the conformable relations with the overlying Amole arkose, and because of the belief that the formation in turn overlies the volcanic series also assigned to the Cretaceous, the red beds are assigned to the Cretaceous. No fossils have been found in the formation in the Tucson Mountains, but poorly preserved forms have been found on the Pyatt Ranch at the northwest end of the Huachuca Range, in what is believed to be the equivalent of this member. Any further study of this problem should include a careful examination of the red beds of the Huachuca Mountains.

Amole arkose.—The Amole arkose formation outcrops chiefly on the western side of the southern part of the range and extends north to Amole Peak, where it is cut off by the granitoid intrusives. The northeastern

part of this belt, however, extends across and around the eastern side of the peak.

The Amole formation is a series of gray to pink coarse-grained arkoses, interbedded with shales and a few limestone beds. It is characterized by rapid and repeated changes in rock type. The arkose beds range from gray to pink and from fine to coarse grained. They weather to either gray or light brown. The interbedded shales are rarely exposed and are usually silvery gray. Few of the limestone units are more than 2 feet thick, and most of them are very thin-bedded. They range from blue-gray to black. They are well exposed on the pediment, where they stand 6 inches to a foot above the interbedded clastics. In unmetamorphosed areas south of the Ajo road, some of the limestone beds are fetid.

The thin-bedded limestones weather to knife-edged ridges between paper-thin shale partings. Under the microscope these partings are seen to be partly disseminated quartz grains and partly bands of more coarsely crystalline calcite. In one section the calcite follows the bedding but has a cross-fibre texture. Part of this calcite is replaced by quartz. Replacement of this type may account for the common cross-fibre veinlets of quartz in the arkose.

Conglomerates are rare. A well-assorted pebble conglomerate is exposed on the synclinal axis on the pediment northeast of Saginaw Hill and another in an arroyo between the Silver Lily mine and Piedmontite Hill.

No satisfactory section of the Amole arkose is exposed in the range, but the section on the west limb of the synclinal ridge south of the Mile Wide mine has been roughly measured. As seen from the north, this section is fairly regular in general outlines but is highly irregular in detail. It is cut by three faults of undetermined, but apparently slight, displacement. Dikes mark each fault.

The section given in Table 3 was measured between the red-bed contact in the arroyo at the tourists' picnic ground near the Ranger Station and the United States Mineral Monument on the hill about a mile northeast. The similarity of some of the limestone beds strongly suggests repetition in the section, but this could not be proven.

TABLE 3.—*Section of Amole arkose at Ranger Station*

Top of hill	Feet
AMOLE ARKOSE FORMATION	
1. Massive fine-grained light-brown arkose	150
2. Brown and silvery-gray shales interbedded with thin-bedded limestones and a few thin arkose beds	500
3. Hornfels and interbedded quartzitic arkose. Metamorphosed by dikes	450
4. Interbedded gray arkoses and light-brown sandy shales with 10-foot zone of thin-bedded limestone beds at top	450
5. Gray arkose beds some of which have such a large proportion of calcareous cement that they weather with the characteristic pitted surface of limestone	100

Top of hill	Feet
6. Gray to brown arkose beds interbedded with brown to black limey shales and thin-bedded limestones. One bed of argillaceous limestone and one of sandy limestone carry a mactra and gastropod fauna	400
7. Gray and pink arkoses which weather light brown, interbedded with sandy shales	150
8. Mostly covered but apparently transition zone of interbedded gray and brick-red arkoses of formation below	75
	2275

Recreation red beds

The Amole arkose is clearly conformable with the underlying red beds into which it grades through a zone about 75 feet thick. The top of the Cretaceous is apparently not exposed in the range but is cut off either by erosion, as in the section measured, or by the great thrust. An angular unconformity or the thrust and the overthrust sheet invariably separates the sediments from the overlying Tertiary volcanics.

Over much of the area the sediments are highly metamorphosed. A typical example is on the road within half a mile of the Mile Wide mine. The arkoses are without distinct bedding, and the shales are massive hornfels. Bedding can only be distinguished by the trend of the ledges. The great Tucson Mountain thrust passes through the mine, and, in the hanging wall block, bedding can be distinguished only in the metamorphosed thin-bedded limestone.

The metamorphosed limestones are altered to a massive epidote-garnet rock with the epidote greatly predominating. Usually the bedding is well preserved in such rock. The shales are changed to hornfels which consists of aggregates of quartz and sericite. Macroscopically, it is light to dark gray, breaks with a conchoidal fracture, and resembles novaculite. "Knots" in the hornfels consist of epidote and pyrite that are formed by a combination of the force of crystallization and replacement, as in one case an epidote-pyrite knot had a greater diameter than the amount of the bending of the bedding around the kernel.

Silicification commonly alters the shales to a fine-grained dense rock similar to hornfels. However, on the small hill north of the Santa Cruz River, arkoses, apparently of Cretaceous age, are silicified to a massive white rock similar to "bull quartz."

The unaltered Amole arkose formation is one of the most easily eroded formations in the range. This weakness has caused part of the sapping at the base of the escarpment on the western side of the mountains and has produced the large areas of pediment where it has been reduced to a gently sloping surface.

In metamorphosed areas, on the contrary, the tough hornfels, quartzites, and garnet-epidote rock are hill formers. The hill just north of the Sagi-

naw stock and the hills west of Golden Gate Mountain are composed of the metamorphosed rocks of this series, and much of the high country on and around Amole Peak also consists largely of these metamorphic rocks.

Fossils are scarce in the formation, but two zones about 300 feet above the base carry a gastropod and pelecypod fauna. Hubert Woods, of the writer's 1927 field class, found the best material in a gulch just east of the Silver Lily prospect. Stoyanow (personal communication) kindly made the following statement concerning the identity and age of the fossils:

"The mactroid lamellibranchs found by you in thin-bedded limestones of the Tucson Mountain area clearly indicate Cretaceous age of the involved rocks. Similar limestone beds are now known at several points between Fish Canyon, south of Greaterville, and the Empire Mountains on the north. I feel rather confident that the strata in question are considerably younger than the *Stoliczkaia* beds in the Patagonia Mountains, farther south, and yet older than the varicolored shales and sandstones with fresh-water invertebrate and reptilian faunas of the Santa Rita Mountains which are to be placed in the later Cretaceous on paleontological evidence. Although the Upper Cretaceous age of the *Mactra*-bearing limestones of the Tucson Mountains is almost certain, their definite placing either in the Colorado or Montana group would not be warranted on the basis of the paleontological material collected up to date."

Similar forms probably identical but poorly preserved are found in the area of folded sediments south of the Ajo road.

Fossil wood occurs in what is apparently a horizon close to this faunal horizon in the folded rocks south of the Ajo road and also in the embayment in the escarpment at the southeast base of Golden Gate Mountain. In the former locality the wood occurs as a single log, lying parallel with the bedding and the surface. It is about 2 feet in diameter and apparently floated into its position. In the Golden Gate locality several logs or fragments of logs occur in a black siliceous bed. On field examination this appeared to be either a silicified coal or peat deposit with fragments of logs in the black matrix.

Through the kindness of Robert S. La Motte, thin sections of these materials were submitted to Lyman H. Daugherty, who states that the logs are "a conifer which is probably an *Araucaria*." The thin sections are too poor to permit closer determination. The cherty matrix material contains

"macerated plant material which Mr. Daugherty considers indeterminate. He found some isolated tracheids of coniferous wood. The mode of occurrence—not the plant remains—suggests somewhat the samples I have seen of the Rhynie cherts from Aberdeenshire, Scotland."

Later, a specimen of the wood was determined by Mr. Daugherty to be "typical *Araucarioxylon*."

According to Knowlton (1919, p. 93), *Araucarioxylon* are known well into the Cretaceous and probably into the Tertiary.

Correlation.—Work on the Cretaceous of southern Arizona hardly warrants precise correlations, but the occurrence of rocks of this age have been noted over a wide area. Wilson (1934, p. 429) has mentioned the occurrence of a Fort Pierre fauna in the Cretaceous of the Empire Mountains. Taliaferro (1933, p. 12-37) described a Cretaceous series to the south, in Sonora, carrying both vertebrate and invertebrate faunas. Darton (1924) classified most of the younger sedimentary rocks in southern Arizona, older than known or supposed Tertiary, as Comanche.

On a visit to the Empire Mountains with Julian Feiss, the writer was shown a section on the edge of Feiss's (1929) thesis area near the Hilltano mine which corresponds almost exactly in stratigraphic sequence to that in the Tucson range, although the thickness of the members apparently differs greatly. The roughly estimated thickness of the members is given in Table 4.

TABLE 4.—*Empire Mountain Cretaceous near Hilltano Mine*

(thicknesses estimated—not measured)

	Feet
Alluvium	
CRETACEOUS	
1. Gray arkose interbedded with gray paper shale and shaly limestone	200
2. Massive gray arkose interbedded with red shale	500
3. Fine-grained red sandstone, minutely jointed. Contains a few thin conglomerates composed chiefly of limestone and quartzite pebbles. There are a few andesitic pebbles in one conglomerate	500–800
4. Basal member a limestone conglomerate with boulders up to a foot in diameter resting disconformably on the Snyders Hill phase of the Permian	30
Permian	

No fossils were found in this section, and it is noteworthy that the Cretaceous volcanic rocks are missing.

The formational units present in the Tucson Mountain section and in the similar section given in Table 4 seem to be represented by lithologically similar formations in both the Huachuca and the Patagonia mountains. Stratigraphic and paleontologic work is needed on this series of rocks before the complicated structure of the Cretaceous areas of southern Arizona can be solved.

LARAMIDE INTRUSIVE ROCKS

General statement.—The Laramide [2] revolution was a time of great disturbance in the Tucson Mountains. During the earlier part of the Cretaceous period, volcanism was active, and a great thickness of lavas and

[2] The writer has used the term Laramide throughout the paper for the revolution between the Cretaceous and the Tertiary. In the Tucson range the age of the revolution is between Amole arkose, Upper Cretaceous, and Tertiary volcanic time. It might seem preferable to designate the revolution as "Late Cretaceous or early Tertiary," but the writer prefers and has used the less awkward term Laramide to designate the age of the revolution and also the age of the slightly later, but accompanying, intrusives.

tuffs accumulated. Relatively quiet conditions ruled later in the Cretaceous. The revolution began with extensive folding and thrust faulting and less extensive northeasterly trending steep-angle faulting. Intrusive rocks invaded this complex structure.

The principal Laramide intrusive body is a stock of granite and quartz monzonite west of Amole Peak. The granite can probably be separated into a granitoid and a porphyritic member, although it is not so separated on Plate 1.

Associated with the stock of approximately the same age is a complex series of latite sills injected along the bedding of the highly disturbed sediments and along the thrust faults. Two small stocks in the southern part of the range are believed to have been intruded at this time.

Amole granite.—The Amole granite occupies the western side of the exposed part of the stock and, although the western border is covered by alluvium, presumably is the central part of the intrusive mass. The area of granite is roughly the shape of an inverted "T," with the top of the "T" trending northwestward parallel to the strike of the invaded sediments. The upright of the "T" is 3½ miles long and trends northeastward, or in the direction of the predominant Laramide steep-angle faults.

The granite is medium grained, with pink felspar crystals about half a centimeter in diameter, quartz, and prominent biotite crystals. On a fresh surface the rock is pink, speckled with black, but unweathered surfaces are rare. On weathered surfaces it is light tan to drab.

The granite occurs in three different topographic positions which govern its appearance on weathered outcrops and the number of outcrops as well. On the western side of the higher part of the range the granite forms bold steep slopes and cliffs. On the lower hills west of the main range aplite dikes, which riddle the granite, resist weathering much better than the granite and to a large extent cover the granite and give the impression that the area is largely aplitic. Careful search must be made in saddles and ravines to determine the presence of the granite. On the flatter slopes of the pediment it is even more difficult to find granite outcrops; they are on the sides of elevations along the dikes.

As seen under the microscope the granite shows variable composition and texture. Predominantly, it is medium to fairly coarse grained, and the more acidic varieties consist of microperthite, quartz, and biotite with granitic texture. In these, plagioclase is scarce and mostly in small interstitial crystals. Hornblende is rare, and sphene and apatite are the most common accessories.

Similar in appearance to the granite in the hand specimen, but different in composition, is a rock with monzonitic texture and containing about

28 per cent acid labradorite, 44 per cent microperthite, and 28 per cent quartz. Biotite is the only important ferromagnesian mineral, and hornblende, sphene, and apatite are in small amounts. These two rocks have been mapped as granite. They probably represent extremes in its composition.

Along its southern margin the granite is clearly intrusive into the Cretaceous sedimentary rocks. The relation is best seen on the eastern side of the hills southwest of Amole Peak that form the westernmost end of the range. There the granite is in contact with the Cretaceous arkose beds, and small apophyses extend into the sedimentary rocks. The border of the intrusive against the arkose grades within a distance of 2 inches from coarse granite to material that resembles pink porcelain. Within the arkose, for half an inch from the contact, minute crystals of biotite have been abundantly developed. Farther from the contact, garnet and epidote occur as veins and bunches in the arkosic and volcanic rocks.

Along its eastern and northern borders the granite is in contact with the granite porphyry, which in turn is in contact with the Amole quartz monzonite. The western border of the granite is covered by the alluvium of the Altar valley fill. The small area of quartz monzonite next to the valley fill suggests that the edge of the granite is not far to the west, in the covered area.

The granite is definitely younger than the Cretaceous sedimentary rocks which it intrudes. It is not in contact with the Tertiary volcanic rocks, but the typical metamorphosed Cretaceous adinoles and hornfels, which are apparently due to metamorphism by the granite, occur in the conglomerate at the base of the Tertiary volcanic rocks on the cliffs south of Safford Peak. This seems to indicate that the granite is Laramide in age.

Granite porphyry.—Along the entire length of the contact of the granite and quartz monzonite, and included with the granite in mapping, is a rock which at first glance resembles the abundant aplites which cut both formations. The width of the belt is variable, ranging from 50 to 200 feet, and in one place reaching 1000 feet. The rock is pink when fresh and on a weathered surface has the light-tan or buff color which characterizes the aplites. However, on examination it is seen to be porphyritic. The groundmass has the texture of aplite, and many of the phenocrysts which are about half a centimeter long are striated feldspars.

In thin section the character of this rock is clear. The phenocrysts are andesine and microperthite in a groundmass consisting of quartz,

microperthite, plagioclase, and biotite. Apatite, sphene, and magnetite are accessories, and there is very little hornblende. The microperthite and quartz of the groundmass appear to replace some of the plagioclase phenocrysts, one of which has an outer zone of orthoclase. Other zoned plagioclase crystals show a range from acid to basic andesine.

The contact of this member of the intrusive series with the quartz monzonite is usually covered by wash and can only be located within 3 to 5 feet. The color contrast with the monzonite is marked, but the granite has practically the same color as the porphyry, and there is some suggestion of a transition. At one point along the contact between the porphyry and the quartz monzonite they mutually interpenetrate. The porphyry has the same resistance to weathering as the aplites, and fragments of it cover the granite to a large extent. All three rocks are cut by the aplites.

Amole quartz monzonite.—To the west and northwest of Amole Peak, and circling the northeasterly portion of the Amole granite, is a belt about a mile wide and 4 miles long of gray quartz monzonite which differs markedly from that found in the granite mass. It is a fine-grained, gray, holocrystalline rock speckled with black crystals of biotite and hornblende. Quartz is difficult to distinguish in the hand specimen.

Under the miscroscope the rock has a distinct monzonitic texture (Grout, 1932, p. 85) with plagioclase and microperthite the principal feldspars. Much of the plagioclase is zoned and throughout has a range from An_{40} to An_{70}. Quartz is abundant in all specimens except the extremely basic ones which grade to diorite. Biotite and hornblende are in about equal amounts. Apatite and sphene are accessory. Throughout the percentage of these constituents ranges considerably, but the writer saw no subdivisions in the field, and in thin section only one sample showed a diorite. Hornblende locally, but rarely, exceeds biotite and in many slides is scarce. Neither mineral is abundant. Quartz and microperthite are scarce in the more basic rocks. Quartz is interstitial. The rock is fresh in most areas, but along its northern border sericitic and chloritic alteration is common. Locally chlorite colors the rock green.

On its northern, southern, and eastern boundaries the quartz monzonite is intrusive into the Cretaceous rocks and on its western side it is in contact with the granite. Along the northern contact the Cretaceous rocks, chiefly the volcanic series, are highly disturbed and altered. Along the eastern border the quartz monzonite mass is in contact with the upper gray portion of the Cretaceous sedimentary rocks and is

usually best exposed where it is in contact with the arkosic members. These contacts commonly appear gradational, but under the microscope it can be seen that the apparent gradation results from the development of minute hornblende crystals in the arkose giving rise to a poikiloblastic texture. The contacts with the finer-grained arkosic hornfels are sharper and show no suggestion of gradation.

The contact relations are best exposed along the southwest tongue of quartz monzonite that extends between the granite and the sedimentary rocks. There the quartz monzonite is frozen against the arkosic rocks, and there also it is difficult to distinguish metamorphosed arkose from fine-grained quartz monzonite.

The quartz monzonite occurs in two different topographic positions—on the slopes of Amole Peak and on the pediment to the west. On the slopes of Amole Peak where denser arkoses and latites crop out above the monzonite, quartz monzonite is generally covered by boulders and is well exposed only where steep hillside ravines cut through the overburden. On hills where there is no denser rock above the quartz monzonite the whole slope is covered by boulders of disintegration which rest on loose arkosic sand. Outcrops in such places are scarce except on the ridges and hillside ravines. On the pediments, outcrops occur on low knolls and along arroyos and are usually disintegrated. The best exposure of the quartz monzonite is that in the steep arroyo heading about 1000 feet west of Amole Peak and extending northward.

The method by which this rock series was developed in the large intrusive mass is a puzzling problem. The contacts between the three rock types are not well exposed, but nowhere was evidence found to indicate that they are of different age. Some process of differentiation seems most likely. The probable sharpness of the contact, which, as noted on a previous page, cannot be placed more closely than within 3 to 5 feet, owing to cover, seems a significant feature in any interpretation. Also, the mutual interpretation of the two major facies in tongues appears to eliminate separate intrusion. The lack of any progressive compositional change in the monzonite and granite and the probable sharpness of the contact between the facies seem to indicate a method of differentiation calling for a sharp separation rather than deposition from the walls inward. Whether this contact is the outer leaf in a magmatic filter press (Bowen, 1928, p. 23) or the interface between two immiscible phases of a liquid magma (Grout, 1932, p. 250) is difficult to determine.

The interstitial occurrence of the quartz in the monzonite and of the plagioclase in the granite suggests that separation took place previous to crystallization of the granite and that mass action determined the order of paragenesis in the two rocks. This paragenetic sequence, how-

ever, would be in accord with either the filter press or the immiscible liquid hypothesis.

Amole latite.—Jenkins and Wilson (1920, p. 13) described "a very peculiar acid intrusive of rhyolitic type" on the top of Amole Peak and noted that it was difficult to distinguish from the invaded sediments. The writer has called this intrusive the Amole latite. It occurs as a series of roughly east-west fingers extending eastward from the quartz monzonite.

The latite shows extreme variety. The commonest is a dense felsitic rock with inconspicuous phenocrysts of quartz and feldspar and abundant xenoliths of arkose and other Cretaceous sedimentary rocks. These xenoliths are commonly altered to aggregates of epidote and in some places tourmaline, giving the rock a spotted green or black appearance. This xenolithic facies of the latite generally ranges through light grays and buffs. Rarely it is red and purple, and where dark purple and loaded with xenoliths it is difficult to distinguish from some of the Cretaceous andesitic breccias.

Elsewhere the rock is coarsely felsitic and so full of fine-grained inclusions that it closely resembles the invaded arkoses. The fine-grained facies usually have a pronounced flow structure and resemble the Cat Mountain rhyolite when inclusions are larger.

In the vicinity of the Amole quartz monzonite the latite increases in grain size and abundance of phenocrysts, but no gradational contacts with the quartz monzonite have been found. The latite in the saddle southeast of Amole Peak has quartz phenocrysts whose shape and distribution give the rock an appearance similar to graphic granite.

Under the microscope the latite is porphyritic, with microperthite, plagioclase, and quartz as phenocrysts. The plagioclase is difficult to determine but appears to be andesine. In a few areas it is absent. The ferromagnesian minerals are completely altered to epidote and chlorite. Both the grain size and the abundance of phenocrysts range widely. Near the quartz monzonite, phenocrysts are so abundant that in the field the rock may appear to be granitoid, although under the microscope the groundmass is distinct. Elsewhere near chilled contacts of the finger-like intrusives there is a well-delevoped flow structure with minute phenocrysts in a cryptocrystalline groundmass. All gradations between the two extremes are found.

Intrusive masses have greatly metamorphosed the invaded rocks and the abundant xenoliths. Of the metamorphic minerals developed, epidote is by far the most common, but brown tourmaline, which is colorless in thin section, black tourmaline, axinite, tremolite, several garnets, jefferisite, and vesuvianite occur abundantly at several localities. The

occurrence of tourmaline and axinite with their "mineralizer elements" seems noteworthy.

The Amole latite is intrusive as sills into the Cretaceous sedimentary and volcanic rocks. Usually finger-like masses follow the bedding of the sedimentary rocks no matter how highly contorted. Elsewhere, as in the saddle on the ridge southeast of Amole Peak, the latite is intrusive along the great thrust zone. Here the sediments at the base of the intrusive have the north-south strike characteristic of the underthrust block, and the sediments above have the east-west strike characteristic of the overthrust block. A similar occurrence of latite intruded sill-like along a thrust is found 5800 feet north of Amole Peak. There, Amole arkose dips under Cretaceous volcanic rocks, and the latite sill follows the contact between the two.

In addition to these larger relationships the latite intricately invades the sedimentary rock along its contacts. In many areas the sediments are so minutely penetrated by latite that it is difficult to decide where to map the boundary. The minute complexity is best seen where the intrusive makes contact with hornfels, as there is commonly a definite contrast between the two rocks. One of the best examples is in the Mile Wide mine where the latite fills a mosaic breccia in hornfels. In contacts between the intrusive and arkose the similarity of the two in the field makes the details difficult or impossible to determine.

Because good contacts between the latite and the quartz monzonite have not been discovered the age relationship of the two is indefinite. Both the spatial relationship and the increase in grain size of the latite toward the larger intrusive mass suggest a genetic relationship. It seems probable that the latite was intruded during an early stage in the invasion of the granitoid rocks, possibly before differentiation was complete, and that the alteration of the latite was in part an effect of the larger intrusive mass.

Pegmatites, aplites, and lamprophyres.—Pegmatites and lamprophyres are relatively rare in the range, although aplites are common. All three are associated with the granitoid intrusives west of Amole Peak.

Only one pegmatite has been found. This is 9500 feet N 79° W of Amole Peak on the south bank of a large arroyo. The patches of pegmatite occur in the granite porphyry. They seem scattered and irregular and grade into the porphyry. None is more than 3 feet in diameter, and they are either cut by or associated with small aplite dikes. The pegmatite consists of quartz and feldspar in crystals up to an inch in diameter. The exposure is probably the erosion remnant of a nearly

flat sheet-like mass only a few inches thick. The feldspar is predominantly microperthite with a small amount of plagioclase.

As noted in the description of the Sedimentary Hills stock, there is an area of silicification with associated feldspar which approaches a pegmatite in character.

Only two lamprophyre dikes have been found. A small one at the head of the gulch west of the Mile Wide mine, near the quartz monzonite contact, ranges from porphyritic to granitoid, and the latter variety has a notably high specific gravity. The dike is poorly exposed, and the granitoid facies can be collected only as small spheroidally weathered pebbles, 1 to 2 inches in diameter. The dike is 1 to 2 feet wide and trends N 55° E. Throughout most of its length it is in the Amole latite, within 200 feet of the quartz monzonite.

The other dike is in the granite, about 2½ miles west of the Mile Wide mine. It is a dense dark felsitic rock and under the microscope appears to be an altered basalt.

Aplite dikes are abundant in the Amole granite and Amole quartz monzonite and extend into the sediments on the ridge between the two intrusives. Those in the sediments are very difficult to distinguish from arkose. The writer has not attempted to map the dikes. In thin section, some of them are composed of quartz, orthoclase as crystals and graphic in the quartz, plagioclase, and muscovite. Others contain quartz with microperthite and magnetite. The aplites are distinctly finer grained than the groundmass of the granite porphyry described as a facies of the granite.

Sedimentary Hills stock.—In the Sedimentary Hills, 1½ miles southwest of Golden Gate Mountain, is a small stock, hour-glass-shaped in plan and about one-tenth of a mile long. The porphyritic rock of the stock is light gray and on some outcrops is stained green by mineralization. Under the microscope it has a phaneritic groundmass with quartz, plagioclase, orthoclase, and altered biotite phenocrysts. Sphene is the most common accessory, and sericite and chlorite are alteration products. The rock is monzonite porphyry.

On the eastern side of the southern end of the intrusive there is an area, about 100 feet in diameter, with disseminated pyrite and chalcopyrite in a siliceous facies of the porphyry. Oxidized grains of the sulphides consist of chalcopyrite surrounded by limonite. The exposure was not sampled but appears to be of low copper content.

In a little saddle on the small hill formed by the intrusive there is a prominent siliceous outcrop. Quartz forms parallel veins in what was probably a sheeted fracture zone in the monzonite, the whole having

a banded structure. Locally the quartz is intergrown with coarse feldspar, and the mass resembles pegmatite. Some of the feldspar is distinctly later than the quartz and cuts across it.

The stock has caused more intense metamorphic effects than would be expected from its size. The arkose is highly indurated and resistant to weathering. The shales is altered to hornfels, and garnet and epidote are found in the calcareous beds. It seems probable that the stock has a much greater subsurface extent than the area of its outcrop.

The intrusive does not come in contact with the Tertiary volcanic rocks, and its age is unknown. Both similarity in composition and metamorphic effects lead to correlation with Laramide intrusives.

Saginaw stock.—A small stock of porphyry intrudes the Cretaceous sediments of Saginaw Hill on the western side of the southern part of the range. It is more resistant to weathering than the intruded sediments.

It is one of the coarsest-grained porphyritic rocks in the range, with feldspar phenocrysts reaching 1½ centimeters in length. The larger crystals are unevenly distributed. Quartz phenocrysts are more abundant but are smaller and much less prominent. No ferromagnesian minerals can be seen in the hand specimens. Microscopically it is similar to the acid dikes of the Silver Lily system but is coarser. Quartz, biotite, and altered orthoclase and plagioclase occur as phenocrysts in a crystalline groundmass. The rock is a latite porphyry.

The most prominent feature of the intrusive mass is its irregular alteration which is associated with mineralization. A quartz vein, carrying copper minerals and also disseminated mineralization, occurs in the stock. According to Allen the disseminated mineralization was drilled by the Calumet and Arizona Copper Company in 1914, but the grade was too low to make commercial ore.

The Cretaceous sediments around the stock are irregularly metamorphosed. Most of the metamorphic effects are to be seen in hornfels and silicified shale, but, locally, garnet and epidote are developed. Just north of the stock a small sill of andesitic intrusive is exposed in a prospect hole. Both the sill and the adjacent sediments are partly altered to garnet.

Exposures of the contact of the porphyry with the surrounding sediments are scarce, but in prospect holes the contact approaches the vertical. Outcrops of the sediments near the contact are fairly abundant, and the area should be studied in detail to determine the structural setting of the intrusive. On a basis of the work already done, it appears to have bored or stoped its way through the sediments, as the dips on all sides

are very uniform, and little evidence of structural disturbance has been seen.

The intrusive is definitely post-Cretaceous, but whether Laramide or Tertiary cannot be determined. The alteration is too intense for petrographic similarity to be a safe guide. On a basis of similarity of associated metamorphism it is tentatively assigned to the Laramide.

TERTIARY EFFUSIVE ROCKS

General statement.—A series of interbedded, acid to intermediate lavas and a tuff of Tertiary age outcrop over half the area of the range. These formations occur in a concordant pile and form the face of the escarpment along the western side and cover most of the eastern or dip slope. The age of the series is established by the occurrence of fossil plants in the tuff and the relations of the series to the Cretaceous sediments.

There are striking similarities in the stratigraphic sequence of the volcanics at the two ends of the range, but, as there is also a distinct difference, the two columnar sections are described separately. Four formations, with a total thickness of 1160 feet, occur in the northern part of the range, and five formations, totaling 2100 feet, in the southern part.

Two intrusive bodies which are regarded as volcanic necks, as well as the source of some of the flows, cut the volcanic rocks. Two other groups of Tertiary intrusives have not been correlated with the flows.

Cat Mountain rhyolite.—The most conspicuous, and one of the most widespread, formations in the Tucson Mountains is the Cat Mountain rhyolite. This is the cliff-forming series of rhyolite flows of which the escarpment on the western side of the range south of Amole Peak is formed. The formation is well exposed on Cat Mountain and Golden Gate Mountain. It also outcrops over wide areas on the dip slope and extends from the vicinity of the Old Yuma mine to the large transverse fault south of the Beehive.

Nearly vertical jointing is pronounced in the formation, and weathering of these joints is one cause of the steepness of the escarpment. On these steep slopes the rock is usually dark reddish brown, except for the basal breccia which is light buff. On more-subdued slopes where weathering is more thorough, the rhyolite assumes many shades of purple, brown, buff, red, blue, and even white. When fresh, however, the flows are red and the mud flows or breccias red or buff.

At the type locality the basal rhyolite includes many fragments of sediments and of the Cretaceous volcanics, some as large as 10 feet in diameter. Near the xenoliths well-developed flow structure parallels the contact. Above this basal phase the rhyolite is more uniform and dense. Near the middle of the cliffs on Cat Mountain there is a poorly defined

bench, formed by a bed similar to the material at the base. This horizon is porous and crowded with foreign material. It resembles a tuff, but microscopic examination suggests that it is probably a mud flow. Above this horizon the rhyolite is dense and massive and characterized by well-developed flow structure.

The thickness of the formation cannot be determined. In the Cat Mountain–Golden Gate Mountain area it measures at least 700 to 800 feet, but the top is not preserved. Farther south, west of the Beehive, the thickness is between 200 and 300 feet. Northeast of Amole Peak both the top and the bottom of the formation are exposed in what appears to be one continuous section, but the structure is too complex to justify an estimate of the thickness.

The stratigraphic relations appear to be well established. It usually rests either on the overthrust mass of Cretaceous sedimentary and volcanic rocks or, where the thrust has been removed by erosion, directly on the truncated edges of the sedimentary rock. Wherever its top is exposed, it is overlain by the Safford tuff or its equivalent.

A distinct relationship between topography and the weathering of this formation can be noted nearly everywhere. At high elevations on steep cliffs the rock is fresh and solid practically to the surface. On more-subdued slopes and on the pediment, however, it becomes friable, due to weathering along the flow structure, and breaks readily into rough and splintery thin-edged slabs.

In thin section there are three varieties of the Cat Mountain rhyolite. The most abundant is rhyolite consisting of small phenocrysts of quartz, usually highly corroded, orthoclase, and biotite. Acid plagioclase and microperthite occur in a few sections. The groundmass ranges from glassy to cryptocrystalline and has a well-developed flow structure. Spherulites occur in some thin sections. Xenoliths, commonly of sandstone or arkose, occur in practically all thin sections.

A tuff caps the rhyolite nearly everywhere. This second variety resembles the flows and is difficult to separate from them in the field. Massive flow-like character distinguishes it from the well-stratified Safford tuff. In the southern part of the range, west of the Beehive, the tuff consists of glass shards in a cryptocrystalline matrix. Farther north it consists of highly corroded phenocrysts of quartz, plagioclase, some of which are euhedral, and orthoclase, in a groundmass of glass shards and cryptocrystalline material.

A third variety is the porous tuff-like bed which forms a bench near the middle of the formation on the cliffs of Cat Mountain. The great quantities of xenoliths obscure the groundmass. In some sections the matrix appears to be the typical glassy groundmass of the flows, but

elsewhere in the same sections it may appear to consist of altered shards. It is apparently best described as a flow breccia, or mud flow.

The distinction of this formation from others in the field is usually simple. The well-developed flow structure, the small phenocrysts of orthoclase and quartz, and the scarcity of biotite make the identification certain on fresh exposures. The tendency of the rock to weather into flat splintery fragments and the wide variety in color distinguish it where weathered. The only doubtful area is the region north of Amole Peak which is riddled with the small Amole latite intrusions. The flow structure in some of these closely simulates the flow structure of the rhyolite.

The source of the Cat Mountain rhyolite is not known. The presence of the prominent flow structure in the flows has been noted at all places where the formation occurs. It seems probable that detailed mapping of the flow structure throughout the range may indicate the source.

Safford tuff.—Throughout the range, tuffs and breccias occur at what is apparently the same stratigraphic horizon in the Tertiary volcanic series. The red to brown cliffs near Safford Peak consist of a coarse stratified volcanic breccia or tuff called the Safford tuff (Pl. 3, fig. 1). Farther south the formation is usually fine grained and well stratified to thin bedded.

The coarse Safford tuff extends about 2000 feet northeast of the peak, where it apparently wedges out. It can be traced 2 miles to the southeast. To the west it occupies the lower part of the dip slope of the cuesta-like escarpment and forms a ridge in the saddle west of the peak.

The total thickness of the tuff cannot be determined, as erosion has removed the upper part. Immediately southwest of Safford Peak the tuff is split by the Upper andesite. Only about 50 feet of the upper portions has been left by erosion, but below the andesite there is about 340 feet. The tuff rests concordantly on the Rillito andesite near Safford Peak, but to the southeast, where the Rillito is missing, it rests directly on the Cretaceous volcanics, with marked angular unconformity.

The formation is distinctly stratified, although poorly sorted. Its materials range from sand to boulders 3 feet in diameter. The boulders consist predominantly of the underlying Rillito andesite. A small quantity of white felsitic material is also present. The tuff ranges through various shades of red and brown and is chalky white where altered. Various shades of yellow and orange on the cliffs are due to lichens.

The tuff is poorly cemented but stands as rounded cliffs. Weathered faces are marked by numerous niches localized in sandy zones. The matrix weathers more readily than the fragments, and in detail the surfaces are bumpy.

About 1¾ miles southeast of the point where the coarse Safford tuff wedges out, and 4 miles southeast of Safford Peak, the fine-grained, thin-bedded variety crops out. At this point the tuff lies between the Cat Mountain rhyolite and the Upper andesite. The tuff, as can be seen in Plate 3, figure 2, is fine grained, thin bedded, and ranges from chalky white through gray to buff. Dips are unusually steep in the area, and the tuff has an apparent thickness of at least 500 feet. Its base appears to grade into the tuffaceous upper part of the rhyolite. Interbedded in the lower part of the tuff are several thin flows of a reddish brown felsitic rock, which, although highly altered, is apparently an acid andesite.

Farther southeast, in the vicinity of Twin Hills and Cat Mountain, the same tuff is exposed in a number of fault blocks. Here the tuff lies between the Cat Mountain rhyolite and the Shorts Ranch andesite. One of the best exposures is in the steep gullies on the south slope of Twin Hills; others are in the headwater basin north of Cat Mountain and just east of the Shorts Ranch house. In these localities the formation appears to range from less than a hundred to several hundred feet in thickness.

Under the microscope the tuff varies greatly in composition and texture. Sections of the finer-grained varieties consist of angular fragments of quartz, felspars, and fine-grained felsitic material apparently from the groundmass of porphyritic rocks. They would be classified as arkoses except for the presence of interbedded finer-grained material containing abundant shards of pumice, besides angular quartz grains.

Zones of thin-bedded varve-like clayey material have been mined for clay, and in these a fossil flora has been found. The best exposure of the fossiliferous beds is in the clay pits west of the intersection of the Mission and Ajo roads. Another exposure is found in the borrow pits just north of the Anklam-Golden Gate Pass road.

A fossil flora from the two localities was submitted to Edward W. Berry, who kindly reported as follows:

> "The material is poor and consists mainly of fragments of several kinds of coriaceous conifers with a few fragments that suggest dicotyledonous leaves but which are too indefinite for certainty. The conifers include detached leaves of *Taxodium*, twigs of *Glyptostrobus*(?), and what appears to be a leaf of *Tumion*. There is also a seed which I do not recognize. Both *Glyptostrobus* and *Taxodium* are not known from horizons older than the Tertiary, and I feel sure that the collection represents a Tertiary horizon. The collection does not permit of any more precise age determination, but ———, I am inclined to think that it is more apt to belong in the later rather than in the earlier half of the Tertiary."

On a basis of this determination and the structural features of the occurrence of the rocks, the writer has placed the entire series of volcanic rocks in the Tertiary. The series consists of a concordant pile resting with marked angular unconformity upon the Cretaceous and overlain by

later basalts with another angular unconformity. It is difficult to see how this sequence of structural events can be dated without placing the volcanic series in the Tertiary. As noted on a later page, the basalts are believed to be Tertiary or Quaternary.

The Safford tuff is not known south of the Ajo Road, but on the western slope of the three small hills between Beehive Peak and Saginaw Hill is a few feet of tuffaceous material which may be its equivalent. This bed rests on the Cat Mountain rhyolite and is overlain by the Diopside andesite. It is associated with arkosic sandstone which locally contains reddish-brown quartzite pebbles up to half a inch in diameter. It is coarser in places, as many large boulders, some of which contain Cretaceous fossils, are loose on the outcrop. The tuff is chalky white and consists essentially of glass shards. It seems to be the equivalent of the Safford tuff but is not shown on the map (Pl. 1), because of its limited distribution and extreme thinness.

Diopside andesite.—In the southern part of the range, about a mile southwest of Beehive Peak, a diopside andesite flow outcrops over about a square mile. The western escarpment of the range is low here, and this flow caps three low hills on the main ridge and crops out on the pediment carved on the dip slope.

The andesite is massive and gray green. It weathers brown, and its rounded exposures contrast sharply with those of the underlying and overlying rhyolites. Weathered surfaces are pitted by the removal of prominent plagioclase crystals, which range in length up to nearly a centimeter. Ferromagnesian phenocrysts are also common.

Under the microscope the rock is seen to consist of acid andesine and diopside phenocrysts in a groundmass of small lath-shaped plagioclase crystals. Magnetite is abundant both in the groundmass and as larger crystals associated with the diopside. Apatite is accessory. Part of the diopside is altered to chlorite.

Locally, this formation is separated from the underlying Cat Mountain rhyolite, on which in most places it appears to lie, by a thin bed which is believed to be the southernmost phase of the Safford tuff.

On hills and steep slopes the formation weathers to smooth, but pitted, rounded surfaces on which desert varnish is common. On the pediment the outcrops are covered with angular products of mechanical disintegration. This material resembles alluvium, but investigation usually proves that it is essentially in place.

Because of complex faulting it is impossible to measure accurately the thickness of the andesite, but it is probably between 200 and 400 feet.

No other formation in the range correlates with the andesite unless it

be the andesite interbedded in the Safford tuff. However, it serves locally as an excellent horizon marker.

Biotite rhyolite.—A massive biotite rhyolite crops out in the southern part of the range, east of the principal escarpment, and forms hills on the dip slope. Its exposures are confined to an area of about 3 square miles.

The fresh rhyolite is red to gray, and, as the name indicates, biotite is abundant. Phenocrysts of quartz and feldspar are so closely packed that in some localities the texture appears to be finely granitoid. Crystals average about 2 millimeters in diameter. Xenoliths of the older formations are very abundant near the base. Inclusions similar to the main mass are abundant throughout and are probably autoclastic. The formation is distinctively massive. No divisions between flows can be seen. About half a mile north of the Beehive, the biotite crystals are oriented, and the rhyolite resembles a schist.

On weathered surfaces the rock ranges from gray through browns and pink to red. Weathering usually produces exfoliation, and on the pediment the thin slabs disintegrate to a mantle rich in biotite.

In thin sections the rock is a porphyry in which the small phenocrysts predominate. Highly corroded quartz is the most abundant constituent. Orthoclase has a fracture pattern resembling microperthite. Plagioclase is much less common. Biotite is abundant and characteristically is bent and torn. The groundmass is finely crystalline and is loaded with xenoliths. The rock is classified as a sodic biotite rhyolite. The thickness of the formation is impossible to determine, but it is at least 600 feet.

The rhyolite rests on the Diopside andesite, as seen west of Beehive Peak. The highest point in the formation is apparently the top of the high hills northeast of Beehive Peak. The great thickness of the formation compared with its slight areal extent suggests that it may be intrusive, but for lack of evidence of intrusive character the writer has listed it with the flows. Stratigraphically the formation belongs between the Diopside andesite and the Shorts Ranch andesite. It may be a siliceous facies of the latter which occurs at its base on the south side of the Twin Hills.

Shorts Ranch andesite.—The youngest member of the Tertiary volcanic series in the southern part of the range is the Shorts Ranch andesite. This formation occurs in a number of downfaulted blocks on the dip slope, south of a line running west from Tucson. The type locality is on Shorts Ranch, 10,000 feet northeast of Cat Mountain.

The rock is one of the most massive of the series of volcanics and when fresh is light purplish gray. It weathers to reddish purple or, less commonly, brown. In the southern part of the range, the andesite has a marked flow structure—visible only on weathered surfaces, which causes it to

weather to a minutely pitted, slabby surface. In the hand specimen the rock has a dense groundmass with abundant phenocrysts of plagioclase, 1 to 5 millimeters in length. Hornblende is common, although some specimens have biotite instead.

West of Tucson and in every area of the andesite north of the Ajo road the formation rests on Safford tuff which establishes its stratigraphic position. The contact can be seen both on the southern side of the hills at the northern end of Shorts Ranch and on Twin Hills. The large area of the formation at the extreme southern end of the range, just north of the Black Hills, is faulted against all the adjoining formations except the alluvium. North of the large fault separating this block of andesite from the other volcanic rocks the sequence in the series is different from that directly west of Tucson, and two formations which are absent farther north occur below the andesite. This causes the formation to appear distinctly higher in the series than at the type locality.

The Shorts Ranch andesite is apparently the "Light colored andesite" of Guild (1905, p. 314), who described it as a biotite or hornblende andesite with acid plagioclase. Microscopically, the rock consists of phenocrysts of medium oligoclase (An_{26}), which show zoning in a few sections and either biotite or hornblende or both. The groundmass is crystalline in nearly all sections. Magnetite occurs both as phenocrysts and in the groundmass. Quartz is rare as phenocrysts but seems to occur in the groundmass. Nowhere can a section of the formation be measured, but it is at least 400 feet thick, and probably much thicker.

This andesite can be correlated with the Upper andesite of the northern end of the range but probably was erupted from a different center.

Tertiary conglomerate.—A slightly consolidated conglomerate is exposed beneath the Rillito andesite around the base of the cliffs on the cuesta-like point west of Safford Peak. This member of the series has been traced 2¼ miles north and one mile south of Safford Peak.

The conglomerate is commonly covered by talus but in some of the steep ravines it is well exposed. It rests with angular unconformity on Cretaceous volcanic rocks and is concordant with the overlying Rillito andesite. Although, in general, poorly consolidated, the top part, immediately under the andesite, is locally highly indurated and cemented. In an open cut a few hundred feet north of Silver Bell pass it is impregnated with gypsum.

The dark Cretaceous andesites dominate the greater part of the large variety of material. Some of the andesite pebbles appear very similar texturally to those in the Cretaceous volcanic series, but the groundmass is bright reddish brown instead of purplish black. In addition to ande-

sitic material the conglomerate contains pebbles of chert, white quartzite, limestone, and, near its southeastern end, abundant pebbles and cobbles of typically banded Cretaceous hornfels. This last constituent is important, as it dates the metamorphism of the Cretaceous sediments as pre-Tertiary. Boulders, especially of the andesites, range up to 2 feet in thickness. Except in a few exposures the conglomerate is neither sorted nor stratified. It reaches a thickness of 50 feet. Locally it appears to be missing.

Rillito andesite.—The Rillito andesite is the prominent cliff-former extending from the cliffs west of Safford Peak to the Silver Bell pass, near the north end of the range.

It ranges from drab to brown on weathered surfaces; it is dense and breaks into angular blocks which retain their angularity and smoothness until a late stage of weathering. The otherwise smooth surfaces are pitted by the weathering of feldspars.

The fresh formation is red to brown and contains numerous, prominent feldspar crystals, most of them about half a centimeter in length. Biotite is commonly seen in hexagonal books about a millimeter in diameter. Flow structure is of variable occurrence but is marked by a rough orientation of feldspars. Locally, sheeting related to flow structure is prominent.

The andesite is unique in that it maintains smooth surfaces on weathered outcrops. Weathering takes place chiefly by spalling. Only when disintegrated does it lose its smoothness.

Under the microscope the rock is porphyritic, with prominent phenocrysts of medium oligoclase (An_{25}), biotite, and, less common diopside in a crystalline groundmass. It is classified as a silicic andesite.

The thickness of the formation is best determined on the cuesta-like point west of Safford Peak, where it is 450 feet. The andesite is separated from the underlying Cretaceous volcanics by a poorly exposed conglomerate. The Rillito andesite is overlain by the Safford tuff, which is in turn overlain by the Upper andesite. The two andesites are most readily distinguished by differences in weathering. The Rillito andesite weathers to a smooth hard surface marked only by pitting caused by the weathered feldspars. This surface contrasts sharply with the minutely rough, porous surface of the overlying Upper andesite.

Upper andesite.—The Upper andesite is the massive cliff-former in the south-facing escarpment between Safford Peak and a point east of Picture Rocks and the Old Yuma mine. It also occurs extensively on the dip slope east of Safford Peak.

Near the peak the formation consists of three different phases. The lowest bed is tuffaceous and usually more nearly resembles a hydrothermally altered phase of the overlying massive material. However, it is locally well stratified, and chalky bedded zones split the more massive chalky material. This tuffaceous facies rests on the Safford tuff and is overlain in turn by the cliff-forming flows. At the base of the flows a chilled vitrophyric zone ranges up to 20 feet in thickness.

The Upper andesite ranges from gray through purplish grays to red when fresh and generally weathers to a reddish brown. It is distinctly porphyritic and has a felsitic groundmass. Plagioclase crystals, up to half a centimeter in length, are either glassy or chalky depending on alteration. Biotite is invariably present and conspicuous.

Weathered surfaces everywhere have a minutely porous appearance. Chalcedony and agate are characteristic of the formation. Some of this material may be of gem grade, especially on the pediment around the east foot of Safford Peak, where it is abundant. It is noteworthy that the agates in this locality are highly colored, and many are delicately banded; but farther east, beyond Picture Rocks, the chalcedony is colorless or white. The various siliceous materials are usually concentrically banded in cavities of irregular shape and size, and quartz crystals coat the inner surface. A few specimens show a repetition of the chalcedony-quartz sequence.

East of Safford Peak the flow is banded and more glassy. In its vitrophyric basal phase the feldspars protrude on weathering. Biotite phenocrysts are less conspicuous than in the main part of the flow, because the glassy groundmass is brown. Associated with the vitrophyric facies is a similar one with a dense brilliant red groundmass, apparently almost as glassy.

Under the microscope the formation proves to be very uniform except for the variety in the groundmass, a feature noted in the field. The main cliff-forming member is porphyritic with phenocrysts of oligoclase (An_{25}), biotite, magnetite, and some diopside. The groundmass is cryptocrystalline. The phenocrysts of the basal glassy facies are the same as those in the cliff-former, but the magnetite and plagioclase contain numerous glass inclusions arranged in an almost graphic pattern. The glass is perlitic and has an index of refraction of 1.492. This index corresponds to that of the average rhyolite obsidian on the curve by George (1924, p. 365).

The total thickness of the Upper andesite is impossible to obtain as it is everywhere exposed to erosion as the topmost member of the series. On the cliffs south of Safford Peak the formation is at least 270 feet thick.

Normally the andesite appears to overlie the Safford tuff. Although immediately southwest of Safford neck the andesite splits the tuff as though by intrusion, it is undoubtedly a flow, as east of Picture Rocks there is a strong development of oxidized flow-top breccia.

The Upper andesite weathers chiefly by granulation to small minutely pitted pellets. Weathering emphasizes the minutely pitted surface which serves to distinguish this formation from the Rillito andesite.

At the northern end of the range, on the hill north of Silver Bell pass, a series of volcanics similar to the Upper andesite is included with it in mapping. In the field it differs from the andesite chiefly in the occurrence of variable amounts of quartz phenocrysts. Under the microscope the rock ranges from a silicic andesite to a rhyolite. It seems probable that this represents a horizon higher in the series of volcanics than any exposed around Safford Peak.

In Silver Bell pass, especially in the northeast side, is a small area of vesicular reddish brown felsite. This consists of a mat of plagioclase needles with phenocrysts of olivine entirely altered to chrysotile and iron oxides. It is an olivine basalt, and it is questionable whether it is part of the Tertiary volcanic series or a downfaulted outlier of the younger basalts.

Southwest of Safford Peak where the andesite splits the Safford tuff, there is a noteworthy occurrence of clastic dikes. The upper portion of the andesite is fractured to a coarse mosaic breccia. The dikes are usually fine grained and in the hand specimen look remarkably like fine-grained aplite. Some have coarse fragments next to the walls. The upper limit of the dikes is the base of the tuff, and they appear to have been filled by the same material. Their clastic texture and their composition as seen in thin section bear out this interpretation. At their lower limit the width of the dikes decreases, and some of them grade into comb-quartz veins.

The similarity between the Upper andesite and the Shorts Ranch andesite, both in the field and microscopically, suggests that they may be correlated. They occupy similar stratigraphic positions, and the only real difference noted is the presence of hornblende in the Shorts Ranch andesite and small amounts of diopside in the Upper andesite. It seems almost certain that the Upper andesite was derived from the Safford volcano. This is indicated by the restriction of the basal tuff to the vicinity of Safford Peak and the changes in texture with distance from the peak. Although probably representing nearly simultaneous differentiates from a common magmatic source, the two andesites must have been erupted from different conduits.

TERTIARY INTRUSIVE ROCKS

Safford dacite neck.—Safford Peak is shown on Plate 4. The dacite mass which is largely confined to the core of the peak is called the Safford dacite. The area of the peak is elliptical. The long axis trends about N 60° E, 2800 feet, and the short axis is 1500 feet. The peak, which towers 600 feet above the surrounding volcanics, is divided into several rounded spires by weathering along prominent fractures.

The dacite is light-gray and its flow structure (Pl. 5, fig. 1) is usually nearly vertical and parallel to its borders. The flow structure is marked on both fresh and weathered surfaces, but there is little tendency for the rock to develop partings like those in most of the rhyolite dikes in the district. The dacite throughout most of its extent is chalky white to light gray when fresh and gray to light drab on weathered surfaces. This is in sharp contrast to the surrounding volcanics. In the chalky white to gray groundmass are fairly abundant phenocrysts of biotite and feldspar about a millimeter long. Three varieties of this rock are recognized in the field—gray unaltered, chalky altered, and glassy.

The fresh dacite is porphyritic, with phenocrysts of quartz, oligoclase, biotite, and small amounts of orthoclase. The groundmass is almost entirely composed of minute spherulites, arranged along the flow structure. The glassy border facies contains the same minerals as phenocrysts but has a groundmass of very clear glass, having an index of refraction of 1.494. One plagioclase phenocryst was found with inclusions of glass in an almost graphic pattern. Although the phenocrysts indicate a dacite, it is noteworthy that the index of the glass is that of an average rhyolite. The chalky dacite is also porphyritic, but the only two minerals recognizable as phenocrysts are quartz and biotite. Secondary quartz encroaches on the spherulitic groundmass. The dacite of the small neck southwest of Safford and a dike extending to the east is very similar to that in the Safford neck, and all have a finely crystalline groundmass.

The dacite pierces the Rillito andesite, the Upper andesite, and the Safford tuff, but the writer was unable to find any well-exposed vertical contacts. The cross-cutting relationships in plan and the vertical flow structure are considered evidence of its vertical cross-cutting habit.

Three kinds of border facies are recognized around the mass. At the southwest end there is a chilled border with 2 feet of glass at the contact with the Safford tuff. At the northern end of the eastern margin there is considerable brecciation along the contact exposed in trenches dug by a prospector. About midway on the southeast side is a short broad glassy apophysis, protruding about 25 feet from the border. Elsewhere the border consists of the normal type of dacite and is marked only by

flow structure. Two dikes of similar dacite from the main mass cut the surrounding volcanics. In the vicinity of the glassy border on the southwest, numerous inclusions of the Rillito andesite are found near the border. These may have come from the adjacent Safford tuff or from the andesite itself. Some inclusions have been attacked and corroded to an hour-glass shape.

Because of its shape, the vertical cross-cutting relations, and the localization of the Safford tuff, as well as changes in the Upper andesite nearby, the Safford dacite mass is regarded as an eroded volcanic neck and the source of these volcanic rocks. No flows corresponding exactly to the dacite have been found, but those at the extreme north end of the range, and included in the Upper andesite, approach it most closely.

A mile southwest of Safford Peak, a much smaller neck rises from the pediment, where it is intrusive into the Cretaceous volcanics. It has practically the same composition as the Safford dacite and is marked by the same type of vertical flow structure. It is noteworthy that a line through these two necks has the trend of the prevailing pre-Tertiary fracture system.

Beehive rhyolite neck.—Beehive Peak, a small but prominent spire, rises above an area of Biotite rhyolite south of the Ajo road. The spire consists of two intersecting dikes of rhyolite porphyry similar to the Biotite rhyolite, but without its characteristic autoclastic structure and xenoliths.

The more continuous dike trends northwesterly and is intersected by a northeasterly dike that cannot be traced away from the hill. The main dike can be traced about 600 feet north and south of the hill. To the south it develops a pronounced flow structure with oriented biotite phenocrysts and is finally lost in the Biotite rhyolite. To the north as far as it has been traced it preserves its massive texture. Farther north, approximately on strike, is an unmetamorphosed but schistose-looking facies of the rhyolite which may be the equivalent of the dike, with a well-developed cleavage parallel to the orientation of the biotite crystals.

The dike rock in the immediate vicinity of the peak is massive and porphyritic and has few xenoliths. Quartz, biotite, and altered feldspar are the common constituents. Much of the feldspar is weathered out, giving the rock a pitted surface. Under the microscope the quartz and biotite are fresh and unaltered, but the feldspars are so highly carbonated that they cannot be recognized. The groundmass is crystalline.

The dike rock south of the peak has a flow structure and, like the Biotite rhyolite, the matrix is crowded with phenocrysts. Under the microscope, quartz is the predominant mineral, with orthoclase and

SAFFORD NECK AND ASSOCIATED VOLCANIC ROCKS

Ti-Safford neck, Tst-Safford tuff, Tra-Rillito andesite, Kv-Cretaceous volcanics.

FIGURE 1. VERTICAL FLOW STRUCTURE OF SAFFORD NECK

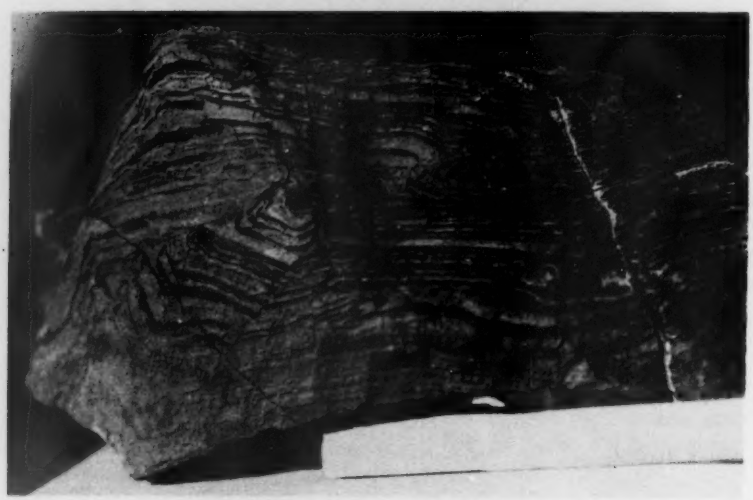

FIGURE 2. TYPICAL SMALL THRUST FAULT
From great thrust fault zone.

STRUCTURAL DETAILS

plagioclase less important. Abundant biotite crystals are bent and torn, apparently by movement after crystallization. This rock is so similar to the Biotite rhyolite that the writer believes that the intersection of the dikes is a volcanic neck and the source of the Biotite rhyolite formation. Unfortunately, however, a debris-covered bench or saddle on the south side of the peak prevents tracing a gradation from the porphyry of the peak to the dike with flow structure on the south slope, and it is difficult to determine on the pediment below whether the dike merges with the Biotite rhyolite. If the gradation can be proved, correlation of the Beehive dikes or dike-shaped neck with the Biotite rhyolite solves the problem of its localization and great thickness.

Silver Lily dikes.—A mile and a half south of Amole Peak a series of nearly parallel dikes extends across the range in the form of an arc concave toward the north. On the western side of the range they are intruded into the Cretaceous sediments above which they rise as discontinuous ridges. In the Cat Mountain rhyolite of the crest and eastern slope the dikes are less prominent topographically. On the crest they tend to form saddles.

The dikes weather to light brown which contrasts sharply with the darker colors of the surrounding rocks. The fresh rock is light buff. There seems to be a definite relationship between width of dike and texture. In the larger dikes, crystals of feldspar and quartz half a centimeter in diameter are common, but in the smaller dikes and in the border facies of the larger ones phenocrysts may be almost lacking. Flow structure parallels the border. The rock cleaves along the flow structure.

Quartz and feldspar can be recognized as phenocrysts in the hand specimen, although in the smaller dikes and chilled borders a hand lens may be needed. As a rule, ferromagnesian minerals are not visible, but biotite is fairly abundant in one dike. The groundmass of the porphyry is felsitic and porcelain-like. Both in texture and in mineral content the dikes are remarkably similar.

Under the microscope the rock is porphyritic, with phenocrysts of corroded quartz rimmed with reaction products. Plagioclase phenocrysts are so altered to sericite and kaolin as to be of uncertain composition. Orthoclase is also altered, and the distribution of the sericite blades in some crystals suggests that it is microperthite with selective replacement of the plagioclase portion. Biotite is usually highly altered. Under low power the groundmass appears to be holocrystalline with indistinct contacts between crystals which extinguish almost uniformly. Under high power the uniformly extinguishing crystals appear to be aggregates and

probably are partly recrystallized spherulites. Probably the rock should be classified as a latite porphyry.

The dikes appear to be the latest rocks in their vicinity. They intrude both the Cretaceous arkoses and red beds as well as the Cretaceous volcanics in the overthrust block. They cut the Cat Mountain rhyolite and the Spherulitic rhyolite, which is regarded as a late intrusive into the rhyolite. On the west the dikes could not be traced into the granitoid intrusives but appeared to stop just short of them. On the east, one of the dikes appears to merge with the Warrens Ranch sill. This relation will be discussed on a later page.

The dikes are from 10 to 30 feet wide. The walls are sharply defined, and the sediments alongside are highly crushed and distorted. The dikes probably occupy faults of small displacement, as the sediments on opposite sides of the dikes have different dips. Veins occur along the walls of some of the dikes. The most prominent vein minerals are siderite and calcite with small amounts of sulphides.

Warrens Ranch latite sill.—At the extreme eastern end of the Silver Lily dike system there is a body of quartz-bearing porphyry similar to the dike rock. This porphyry outcrops over an elliptical area, 1½ miles by half a mile in extent.

The porphyry is so similar to that in the dikes that a description is largely repetition. It consists essentially of quartz, plagioclase, altered biotite, and orthoclase, in a groundmass very similar to that in the dikes.

The porphyry, which is probably a latite, rests on a narrow belt of Safford tuff on the south and is apparently in fault contact with the Cat Mountain rhyolite on the north, although the actual fault has not been observed. One of the dikes seems to disappear in the porphyry mass. It is probable that the two intrusions are connected and that the dike served as a feeder to the sill.

At the same stratigraphic horizon on Twin Hills, about a mile to the south, a quartz-bearing facies at the base of the Shorts Ranch andesite was noted. It is possible that this is also Warrens Ranch latite.

Spherulitic rhyolite.—In the southern part of the range a rhyolite characterized by large and abundant spherulites is called the Spherulitic rhyolite. This is almost entirely in the zone of the thrust and unconformity at the base of the escarpment and is limited to that horizon. It forms outliers on the pediment south of the Ajo road. The northernmost outcrop of the rhyolite is on the escarpment just north of Golden Gate Mountain.

The exfoliated outcrops are light brownish white or buff, bold, and rounded. They contrast strongly with the darker reddish brown of the

escarpment-forming Cat Mountain rhyolite. The outcrops are discontinuous and reach a maximum length of about 1000 feet.

Three textural varieties can be recognized. The most abundant is a felsitic rock having flow structure emphasized by silicification along the flow lines. The second variety has a similar flow structure, but the intervening space between the flow lines and the silica veinlets is closely packed with round spherulites, averaging a quarter of an inch in diameter or less, but reaching a maximum of half an inch. The third is fine grained and shows no flow structure.

The first two varieties make up the main mass, and the third occurs as minute dikelets up to 1 or 2 inches thick, cutting the others and cementing breccia zones.

In thin section the first two varieties are similar and consist of alternating bands of spherulites arranged like beads on a string in a cryptocrystalline groundmass. Quartz phenocrysts are almost all deeply corroded. Accompanying the quartz are phenocrysts of silicic plagioclase and microperthite. The rock is apparently a very sodic rhyolite.

The third variety, however, is a fine-grained intergrowth of quartz and orthoclase. Patches of limonite and a carbonate as well as shadows in the cryptocrystalline material suggest replaced or devitrified glass shards. This seems unusual in dikelets. Possibly these bodies are clastic dikes of pumice.

The relations of the formation are not clear. It invariably occurs in the complex zone caused by the thrust and unconformity at the base of the escarpment. It protrudes through the Cat Mountain rhyolite at places on the eastern slope. Its flow structure is usually very steep, and its contacts appear to be nearly concordant with the overlying Cat Mountain rhyolite, except locally, where it appears to invade that formation. Because of these observations the writer is inclined to believe that the formation intrudes the Cat Mountain rhyolite as well as the unconformity and thrust zone. However, this interpretation is questionable because the formation occurs mostly at the zone of maximum talus accumulation along the base of the escarpment.

TERTIARY (?) LAKE BEDS

The southeasternmost outcrops in the range, other than the basalts of the Black Hills area, are composed of limestone of unique character. The limestone occurs in beds about a foot thick interbedded with slightly calcareous shale. Included in the series are one or more beds of white pumiceous ash and gypsum. The fresh limestone is usually yellowish to brownish gray, and weathers dark gray. The shale beds are buff.

These beds form a syncline with maximum dips of 20 degrees. On the

west the limestone basin is bounded by the Shorts Ranch andesite, but the contact is largely covered, and the relief is so low that the dip of the contact is indeterminable. Along its northern border the limestone series apparently rests unconformably on the Biotite rhyolite. Near the center of the limestone area is an abandoned vertical shaft in which the alternating beds of limestone and shale are well exposed, but, as the shaft was unsafe to enter, more could not be seen. On the dump at the shaft is a considerable amount of Shorts Ranch andesite, and it appears that this formation underlies the sediments.

Fossils of two types have been found in the series. In the shale fragments lying on the dump of the shaft are abundant ostracods. Through the kindness of John B. Reeside, Jr., these were submitted to several specialists, but he writes, "No one will admit ability to determine the simple cyprid ostracods present." As these simple unornamented forms extend throughout the geologic column they are of no use in age determination. Lying loose on the surface, a slab of limestone was found, made up of what appeared to be concentric algal structures about an inch in diameter. These have not been submitted to an algal specialist.

The determination of the age and relations of these beds is of considerable interest and importance. Their apparent position on the Tertiary volcanics with an erosion surface beneath, the presence of the interbedded pumiceous material, and their lack of resemblance to any horizons in the older formations lead the writer to the belief that they represent late Tertiary lake beds deposited on a flat erosional surface or pediment formed during the Tertiary. An alternative interpretation is that they are of pre-Tertiary age, in which case the beds are underlain by a thrust fault. The beds should be more carefully studied and the shaft examined to determine which interpretation is correct.

TERTIARY OR QUATERNARY BASALTS

Basalts occupy two large and one small area, Tumamoc Hill, the Black Hills, and an area in Silver Bell Pass. The spur of the mountains extending to the city of Tucson is comprised of two parts, the eastern known as "A" Mountain or Sentinel Hill, and the western as Tumamoc Hill. These two hills have been studied and mapped in detail by Tolman (1909, p. 67-82), whose report includes a chapter on the petrography by Guild which revises Guild's earlier paper on the petrography of the range. At the extreme southern end of the range are the Black Hills and a number of small hills in the vicinity. Near the northern end of the range in Silver Bell Pass, basalt outcrops over a small area and has been included in mapping with the Upper andesite. Whether this outcrop should be included with the Tertiary or Quaternary basalts is open to question.

In the Tumamoc Hill area, Tolman and Guild have recognized olivine basalt, plagioclase basalt with very large phenocrysts, amygdaloidal basalt, and rhyolite tuff. These lie in an essentially horizontal pile in which Tolman has recognized intrusives. The writer did not attempt to check Tolman's detail except along the contact with the underlying andesite.

The Black Hills are composed of a series of basalt flows similar to those in the Tumamoc Hill area. The predominant olivine basalts have vesicular tops, some of which are oxidized. Included in the series is the rare quartz basalt described by Guild. The flows dip gently to the north. The base of the series is not exposed, but an unconsolidated conglomerate at the base of the slope, north of the Black Hills triangulation station, composed of typical Cretaceous arkose, probably represents the underlying formation.

An intrusive mass of very coarse porphyritic plagioclase basalt similar to that mapped by Tolman in the Tumamoc Hill area extends southwest from the triangulation station. The intrusive is about 500 feet in width and is exposed for about 2500 feet. The main axis of the intrusive is parallel to the trend of the hills. On strike with it is an outcrop of coarse porphyritic basalt on the basaltic pediment at San Xavier Mission. The trend of the intrusive, parallel with the axis of the hills and with the trend of the transverse faults in the Tertiary rocks to the north, suggests that the Black Hills basalts are the result of a fissure eruption. The occurrence of the Tumamoc Hill intrusive B_1 and B_2, mapped by Tolman on a line of similar trend and, in turn, on the trend of a large fault supports this idea. The writer has not attempted to remap Tolman's intrusives, but the evidence for intrusion in the Black Hills seems conclusive. The coarsely porphyritic plagioclase basalt is definitely cross-cutting. It has a step flow structure. The highly oxidized character of the intrusive probably indicates that the exposed rock solidified near the surface.

The age of the Tucson Mountain basalts cannot now be established, although their position in the chronological sequence is clear. The basaltic eruptions are younger than the tilting and faulting of the range, because they rest horizontally on the tilted flows in the Tumamoc Hill area. The eruption, however, took place soon after the tilting, as the surface of the underlying andesites was relatively rough, and erosion had apparently not reached the Cat Mountain rhyolite, which forms most of the dip slope at present. This is shown by the abundant Shorts Ranch andesite and lack of Cat Mountain rhyolite in the conglomerate interbedded in the basalts. The occurrence of pediments carved on the basalts at the San Xavier Mission and at the southwest end of the Black Hills indicates

that they cannot be correlated with the Recent basalts of the Papago country described by Bryan. A coarsely porphyritic basalt, similar to those already described, occurs along the Southern Pacific Railroad, at the foot of the Rincon Range, between Pantano and Vail. This basalt is associated with steeply tilted valley fill. It seems probable that detailed work in that area may establish the age of the basaltic eruptions. For the present the writer is inclined to estimate the age as Tertiary or Quaternary.

QUATERNARY ALLUVIUM

Gilbert described the Basin and Range mountains as "island mountains buried in a sea of alluvium." This applies perfectly to the Tucson Mountains. There are two distinct types of alluvium. On and forming the bajada slopes is coarse alluvium or slope wash. Fine silt is found on the flood plains of the axial streams of the intermontane troughs and deposited on the low-gradient slopes by sheet floods after sorting from the coarser materials above.

The slope wash is poorly sorted and contains material ranging from large boulders to fine silt. Single beds in an outcrop may be sorted and contain boulders with a matrix of sand, and adjoining beds may be made up of pebbles and sand. This amount of sorting emphasizes the stratification, which is a striking feature of the deposits. The individual fragments in the wash are usually angular to subangular. Most of them are of local origin. Exceptional fragments of remote origin will be discussed in following pages.

The silt of the flood plains and the lower alluvial slopes contrasts sharply with the slope wash. It is fine and, where exposed along the Santa Cruz River, is well sorted and in beds usually 6 inches to a foot thick, separated by distinct stratification. The trenching of the Santa Cruz River channel took place about 1900, and before that time the flood plain was the site of rather rapid deposition. At the crossing east of San Xavier Mission (just off Plate 1), the silt is well exposed in the banks of the river. There, in the silt, about 3 feet below the surface of the pre-trenching flood plain, a tin can was found buried in the alluvium. At a lower horizon are cattle bones, and at a still lower one is a zone of pottery shards.

Tolman has discussed the relation of slope wash and playa deposits, and the relations illustrated seem to apply equally well to floodplain deposits. In both, the coarse slope wash and fine silt seem to interfinger owing to the migration of the zone of overlap with climatic variations.

The relationship of the fine-grained silt of the lower, flatter slopes to the slope wash is somewhat different. West of the extreme northern end of the range the flat slope toward the Altar River is mantled with fine-grained silt. The transition from coarse slope wash to the fine silt is abrupt. The cause is not clear, but it seems as if the dusty, silt-covered slope, on which "dust devils" are common, either contains a veneer of wind-blown dust or it is the site of remarkable sorting by sheet floods.

Usually the slope wash material is of local derivation, but there are several noteworthy exceptions, two of which have already been mentioned in the discussion of physiography. The alluvium derived from Amole Peak and now occurring 3½ miles farther north has been interpreted as evidence of a former deeper burial of the range. This interpretation is strengthened by a similar alluvium, composed of the Shorts Ranch andesite, north of the Twin Hills. At the southwestern end of the Black Hills the alluvium is predominantly granitic and apparently derived from the Sierrita Mountains. Similarly, much of the alluvium on which the city of Tucson rests is gneissic and was derived form the Santa Catalina or Rincon Mountains. Thus the mountains on the opposite sides of the Tucson range have supplied alluvium which has tended to raise the base level along its sides and assist in its burial.

Cementing much of the alluvium is the widespread caliche. The two are so closely associated that there is a tendency for local inhabitants to call alluvium "caliche". The commonest type of caliche occurs in the slope wash, as beds of calcareous cement which are usually hard and dense at the top, with a distinct crustification. It becomes progressively more porous until it disappears into the alluvium below. Caliche beds are usually from a few inches to several feet thick, and several beds may occur in a single exposure in a 10-foot arroyo bank.

Caliche is concentrated in many places at the contact of alluvium with bedrock and cements the basal slope wash into a solid rubble breccia. The commonest occurrence of caliche cementing rubble breccia is at the base of limestone and basaltic hills. As noted on an earlier page, this cemented breccia is believed to act as a reinforcement at the base of limestone hills and to retard erosion.

In many parts of the district the superficial portion of the bedrock is a highly shattered zone in many places thoroughly cemented by caliche. In some places the cementation is so strong that arroyos have cut laterally into the uncemented shattered bedrock, leaving an overhanging cliff of cemented bedrock above a niche several times as deep as its height. Alluvium, more or less cemented by caliche, may rest on the roof of the niche.

STRUCTURE

INTRODUCTION

The Tucson Mountains are a unit of the Basin and Range system and exhibit many of the structural and physiographic features typical of tilted blocks. They differ from the ranges farther north in Nevada and Utah in the much greater denudation and in lack of evidence of recent faulting. Erosion has reduced the range to a small part of its former volume. It has obliterated all of the marginal parts of the mountains, leaving pediments. Thus, all evidence of the origin of the tilted block structure must be gained from the interior of the range.

Structurally the range consists of three principal parts: (1) A complex basement block of Cretaceous and older rocks, on the peneplaned edges of which lie; (2) a series of tilted Tertiary volcanic rocks which give the range its tilted block character; and (3) a series of basalts that lie almost horizontally on the tilted Tertiary volcanic rocks. Alluvium of the valley fill laps up on the edges of the bedrock formations.

The basement of Cretaceous and older rocks consists of a syncline of Cretaceous rocks which has been overridden from the west by Paleozoic sediments and Cretaceous volcanic rocks on a great thrust fault. Intruded into this folded and faulted structure is a large stock in the west-central part of the range, and many smaller intrusives. Erosion reduced this basement block to a peneplain and almost entirely removed the overthrust block before the eruption of the Tertiary volcanic rocks.

The Tertiary volcanic rocks rest on the older rocks with marked angular unconformity. The volcanic pile dips eastward and is highly disturbed by steep-angle tensional faulting, causing parts of the area to assume the pattern of a fault mosaic.

After the tilting and faulting and after erosion of the volcanics to a rugged surface, basaltic flows were poured out on parts of the edge of the range. These basaltic flows exist in two areas and have been faulted and tilted at low angles, probably by renewed movement along the older structural lines. Since this faulting the range has been modified only by erosion and deposition.

PRE-TERTIARY BASEMENT STRUCTURE

The basement of pre-Tertiary rocks is exposed on the pediments along the western side of the range and to a lesser extent along the northeastern side. The rocks of this basement series range in age from pre-Cambrian to Cretaceous. The major structure is a large syncline of Cretaceous sediments which is overridden by Paleozoic sediments and Cretaceous volcanics on a great thrust fault.

The thrust is obscure because its plane closely parallels the pre-Tertiary

lava erosion surface, and nearly all of the overthrust block was removed by this erosion. Where exposed by the erosion of the overlying Tertiary lavas the thrust is marked by the presence of numerous klippen of Carboniferous limestone resting on the Cretaceous rocks.

Along the western escarpment at the base of the cliff-forming lavas and at the top of the slope-forming Cretaceous sediments is a conspicuous belt of boulder-like masses of Carboniferous limestone as much as 100 feet in diameter and less conspicuous Cretaceous volcanic rocks. This belt is nowhere of great thickness and is discontinuous. Its contact with the underlying sediments is usually covered and obscure, but it clearly rests on the Cretaceous Amole arkose formation. Southward, this belt of Carboniferous limestone and Cretaceous volcanic masses can be traced to a point on the low escarpment opposite Saginaw Hill. Northward, the belt widens and forks where the escarpment ends on the southern shoulder of Amole Peak. The eastern fork follows the base of the rhyolites around the area of Cretaceous on the eastern side of the range, and there the Cretaceous is spotted with patches of Carboniferous limestone.

These masses of Carboniferous limestone clearly rest on the Cretaceous, and the Cretaceous volcanic rocks also rest on the Amole arkose in the same belt. The western branch of Carboniferous outcrops is greatly obscured by the intrusions around Amole Peak, but in that area also the Carboniferous clearly rests on the Cretaceous sediments.

These relations have been interpreted as a great thrust fault, which practically parallels the surface of pre-Tertiary erosion, and whose overthrust block was largely removed by this erosion.

Because of the location of the thrust plane, immediately below the lava escarpment, it is usually covered by talus or slide rock, but it is well exposed in the rear of the Abercrombie Ranch house, 9000 feet N 40°E from Amole Peak. This house is built on the dump of a prospect, and in the cut at the mouth of the tunnel the fault is well exposed. There it consists of a shear zone, underlain by altered Cretaceous volcanic rocks and overlain by the massive Carboniferous limestone. The shear zone is intensely mineralized with specular hematite, and has some copper minerals.

In the vicinity of Amole Peak the thrust is intruded by Amole latite sills. Branching sills extend up along the bedding of the sediments in the overthrust block. The branch of the fault that separates the Cretaceous volcanic rocks and sediments on the north side of the peak is also intruded by a sill.

Associated with the thrust zone in the thin-bedded Cretaceous sediments are minute thrusts, such as that shown in Plate 5, figure 2. All these specimens were loose on the surface or in slope wash; this is unfortunate, because if in place they would give valuable information on the major

movement. Further search for such minor structures may well prove profitable.

Details of the structure of the main thrust zone are not clear. The distribution of outcrops along the covered base of the western escarpment suggests that the belt of overthrust masses is a jumble and that the thrusting took place under very light load. In this jumble zone, and associated with the Paleozoic limestone blocks, are many rubble breccias composed chiefly of limestone and subordinate quartzite fragments. Recently, B. S. Butler has discovered on the eastern side of the range an outcrop which, instead of having the customary breccia appearance, is clearly a conglomerate with rounded cobbles of quartzite and limestone. This conglomerate occurs between one of the larger limestone blocks and the Cretaceous formations, suggesting that the fault may have extended out onto the surface. Elsewhere on the eastern side of the range the limestone blocks are tilted and appear to be both resting on and overlain by the Cretaceous. This would suggest the imbricate type of thrusting. A third possibility is that the limestone blocks are remnants of a great sheet which has been practically entirely removed by erosion, leaving numerous small klippen on the thrust. In this case the occurrence of the limestone blocks, partly in and partly on the Cretaceous, would be explained by later faulting instead of by imbricate structure. A detailed study of one of the areas of disconnected limestone blocks on the eastern side of the range to determine whether they fit into any definite structural pattern will go far toward solving the problem.

Inasmuch as the thrust plane was rotated by the Basin and Range tilting the direction of the thrust is not clear. B. S. Butler called the writer's attention to the fact that restoration of the rotation would produce a horizontal attitude or even a western dip similar to that of other thrusts in the Rocky Mountain region. In addition to this concept a further test has been applied by studying the folding in the Cretaceous sediments on the pediment south of the Ajo road. A section across these sediments shows that the folds are distinctly asymmetrical if viewed in their present attitude and that the steep limbs of the anticlines are on the eastern side. However, if the folds are rotated approximately 20 degrees they lose their asymmetry. The dips of the lavas in that part of the range are comparatively flat and certainly well under 20 degrees. It is believed that this may be taken as additional or confirmatory evidence that the forces causing the thrusting and folding were from the west.

The interpretation of a thrust from the west is also in accordance with the interpretation needed to explain the occurrence of the isolated limestone hills on the western side of the range, Snyder's Hill and the Picacho de la Calera. If these masses of limestone are considered to be the western

limb of the syncline, it is difficult to explain why they are isolated and not a continuous ridge. Minor thrusts are known in the Picacho and the hill to the north. Around the north base and on the dump of an inaccessible shaft at the northwest corner of Snyder's Hill is a small quantity of reddish-brown sandstone which may be Cretaceous but which was not separated in mapping. All these features point toward the probability of major thrusting in or under these hills, and the interpretation of their major structure as klippen on the Cretaceous seems logical.

The only difficulty in this interpretation lies in the indefinite location of the branch of the big thrust between the Cretaceous sediments and the volcanics east of Amole Peak. From the evidence available, the writer has been inclined to think that this fault has an eastward dip, probably steep, as shown on section B of Plate 1. This may be in error; it requires further study.

Cutting the folded and faulted rocks is a series of intrusives which tend to complicate the picture. The sills in the vicinity of Amole Peak have already been mentioned. These intrude the great thrust, the bedding of the sediments, and locally cross-cut the structure. The sediments are highly metamorphosed by the intrusives, and many are difficult to distinguish from the intrusives. The larger stock appears to cut across and obliterate the northern end of the thrust.

TERTIARY VOLCANIC STRUCTURE

The Tertiary volcanic series rests with marked angular unconformity on the peneplaned surface of the complex just described. These rocks, as noted on a previous page, are composed of formations most of which are thick. The Safford tuff near the middle of the series makes an excellent marker horizon. Near the southern end of the range the Diopside andesite is an excellent marker. Near these marker horizons and at the contacts between the thicker formations, much more structure can be seen than in the thick massive horizons. Undoubtedly, much structure has been missed in the massive horizons, and it will be necessary for the Cat Mountain rhyolite to be subdivided before this structure can be deciphered.

The lavas dip easterly from 10 to 20 degrees. This tilted lava cap makes the tilted block character of the range so apparent. The lavas are complexly block faulted by two systems of faults. One set strikes from east-west to northwest and the other from north-south to northeast. By these two groups of faults the Tertiary volcanic rocks are broken into a mosaic, the best example of which is west of Beehive Peak. This is brought out by the presence of the thin Diopside andesite. Both in this mosaic and elsewhere the northerly system of faults appears to be the older, as they terminate against the easterly transverse faults.

The displacement on the easterly transverse faults is almost uniformly south side down. In most places the dip is steep, approaching the vertical. The direction of displacement on the northerly faults is less regular. The amount of displacement on the individual faults is usually difficult to determine exactly, but, as Section G-G′ of Plate 1 shows, the total throw from the north to the south end of the range on the transverse easterly faults is in the neighborhood of a mile.

The character of the fault zones is also variable. Some are clean-cut breaks, with a few feet of rubble and gouge filling. Others, such as the one about 3 miles north of the Black Hills, are a complex zone about 1000 feet wide, with jumbled large blocks of all the formations involved in the fault.

Several of the larger transverse faults should apparently be classed as obsequent. The Twin Hills are on the down-throw side of the fault at their northern base and rise high above the low but upthrown northern side. The occurrence of such an obsequent scarp gives some idea of the extent to which the range has been stripped by erosion since faulting.

All the structures appear to be of the steep-angle tensional type. Only at one point in the range is there any positive evidence of thrusting in the Tertiary rocks.

Just west of the intersection of the Ajo road and the road to San Xavier Mission is a small conical hill, in the eastern side of which there are several clay pits. The hill is composed of Shorts Ranch andesite resting on the Safford tuff. The structure of this hill is somewhat more complex than appears at first glance. Separating the tuff and the Shorts Ranch andesite is a brown andesitic rock that appears to be the member that usually occurs between the Cat Mountain rhyolite and the tuff. In the largest clay pit on the eastern side of the hill the exposed brown andesite rests on the tuff. The base of the andesite is brecciated, and the underlying tuff contains prominent fractures or faults. The brecciated base of the andesite and the fractures truncate the bedding of the tuff. In a smaller clay pit to the south a small thrust fault can be seen in the tuff. It would appear that this is an area of minor Tertiary thrusting, but further work is needed to clarify the relations. A single block of limestone in the zone of brecciation suggests that the brown andesite is Cretaceous.

TERTIARY (?) LAKE-BED STRUCTURE

Resting on the Tertiary volcanics at the southeast end of the range north of the Black Hills is a series of sediments of undetermined age. These sediments have interbedded tuffaceous material and occur as a very flat syncline. They are faulted against volcanics along the western border and rest on them on the north. In the center of the area of sediments there

is a shaft with volcanic material on the dump. The sediments are either post-volcanic and rest unconformably on the volcanics, or pre-volcanic and have been thrust on the volcanics. The fossils so far discovered in the series do not allow a determination of the age, and no definite choice between the two alternatives can be made. The writer is inclined to believe that the sediments rest unconformably on the volcanics.

BASALTIC MESA STRUCTURE

Throughout most of the range the block-faulted Tertiary volcanics are the youngest rocks. However, in a few limited areas there are younger formations which may have been much more extensive in the past and whose interpretation reveals important chapters in the structural history.

The most important of these are the basaltic rocks which comprise "A" Mountain, or Sentinel Hill, and the larger mass of Tumamoc Hill to the west, as well as the larger Black Hills and small knobs in the vicinity of San Xavier Mission.

Structurally the "A" Mountain–Tumamoc Hill mass of basalts is a mesa of basaltic flows situated at the intersection of a number of faults in the underlying Tertiary volcanics. The basalts overlie the Shorts Ranch andesite, and the contact between them appears to be an erosional surface with a relatively steep slope to the north. The surface is not well exposed, but its trend can be determined by the distribution of outcrops. The flat-lying basalts seem to have been poured out on a relatively rugged surface of the underlying volcanics after they had been tilted. It is noteworthy that the bed of alluvium mapped by Tolman as interbedded in the basalts is very similar to modern slope wash, well stratified but poorly sorted. The Cat Mountain rhyolite was not recognized in the fragments, but fragments of the Shorts Ranch andesite were abundant. Apparently the basaltic extrusions took place so soon after the tilting of the volcanics that the Cat Mountain rhyolite was not extensively exposed as it is today. As noted in the description of the Black Hills basalts, there is also a basaltic intrusive cutting the flows in that area.

Tolman found evidence of post-basaltic faulting, but in general the basalts in the "A" Mountain–Tumamoc Hill area are nearly flat lying and appear to be relatively undisturbed. Likewise, in the Black Hills the basaltic lavas are nearly flat, but on Martinez Hill, across the Santa Cruz River from San Xavier Mission, the basalts are steeply tilted. The concentration of disturbance in the basalts along the borders of the range seems to indicate renewed movement along the main mountain block faults in post-basaltic time. Too little work has been done in the basaltic areas of the range, and it appears necessary to study these further.

REGIONAL STRUCTURE

Since the early controversy between Gilbert (1874, 1875; 1928) and Spurr (1901) on the origin of the Basin Range structure, much more data have been accumulated, and the ideas of many workers [3] have tended to crystallize. It seems worth while to see what light this study of a typical Basin Range mountain throws on the general problem of structure and origin.

In review it may be said that the Tucson range consists of three principal structural units. The older rocks form the pre-Laramide block of folded, thrust-faulted, and intruded sedimentary and volcanic rocks which was eroded to a surface approaching a peneplain. This was buried beneath a thick series of Tertiary volcanic rocks, which were tilted more or less uniformly toward the east, intensely faulted by at least two groups of normal faults, and eroded to a rugged surface. On this surface rests a series of basaltic flows, mostly flat lying, but locally disturbed.

Geographically, the Tucson Mountains lie between the Catalina-Rincon range on the east and the Rosruge range on the west. Directly north and on the strike of the volcanic rocks of the Tucson Mountains is the Tortolita range; to the south are the small hills collectively known as Mineral Hill. Structurally, the lava-capped tilted block of the Tucson range dips under the broad alluvium-filled trough of the Santa Cruz and toward the crystalline mass of the Catalina Mountains. The Rosruge range, according to Darton's map of Arizona, consists predominantly of "Tertiary and Cretaceous volcanics", covered along the eastern border of the range by Quaternary volcanics. On a brief reconnaissance trip to the range the writer found that the older volcanics were very similar to the Tertiary volcanics in the Tucson range and that on the eastern dip slope of these were superimposed an essentially horizontal series of basalts.

It has been pointed out on a previous page that the structural disturbance in the Tertiary volcanic rocks consists predominantly of normal faults. These faults in general fit into two groups, those with a north-south to northeasterly strike and those with an east-west to northwesterly strike. In the transverse east-west system the down throw is almost entirely on the southern side.

This south-side-down displacement on the Tertiary faults seems to fit perfectly into the regional pattern of faulting in which the Tortolita range to the north has been uplifted. The rocks of the Tucson Mountains have been gradually stepped down from this uplift to the north. If there is a trough between the north end of the Tucson Mountains and the southern

[3] Because of lack of library facilities the writer has been unable to completely review the Basin Range literature but feels compelled to mention the effect of the work of Davis on the physiographic and Louderback on the structural phase of the problem.

end of the Tortolita range it is narrow, as is shown by the hill composed of Cretaceous arkose midway between the ranges north of the Santa Cruz River.

The sequence of eastward-dipping Rosruge lavas and Tucson volcanic rocks dipping eastward toward the crystalline Catalinas very strongly suggests that the ranges consist of tilted fault blocks, especially if it is considered that block faulting is the predominant type of Tertiary structural deformation in the Tucson Mountains. The valuable generalization usually credited to Van Hise that "minor structure portrays the major structure" is still one of the best guides available in such problems. The great reduction of the ranges by erosion and the filling of the intermontane troughs by alluvium eliminates the possibility of seeing the actual faults.

The occurrence of the basalts on the borders of the Tucson and Rosruge ranges suggests that the basalts may have been intruded along the border faults. This idea is supported by the occurrence of the "A" Mountain-Tumamoc Hill basalts at the intersection of a series of faults near the border of the range. Renewed faulting along the hypothetical border faults in post-basalt time is believed to be responsible for the steep dip of the basalts on Martinez Hill east of San Xavier Mission.

Further progress on the Basin Range problem in this area of great erosion and burial will probably be made most rapidly if the shape of the suballuvial surface can be determined. Probably this could be determined by reflection seismographic methods. It seems advisable to await a more-detailed study of the Tertiary faulting, which must be based on a better knowledge of the stratigraphy of the Cat Mountain rhyolite, before any analysis of the structure of the range and its bearing on relationships to adjoining ranges is made.

GROUND WATER

Since the beginning of the study of the range in 1925, homesteading has increased until nearly all of the eastern slope has been taken up. The greatest problem facing these homesteaders is the development of an adequate supply of water.

The usual method of developing water is by means of a drilled well, and on many of the properties wells have been drilled. So far as the writer has been able to determine, all the wells have yielded enough water for household purposes, but, although no measurements have been made, the quantity of water appears to be distinctly variable and definitely related to the geologic structure.

At the northern end of the range it has been noted that the volcanics have at their base a somewhat discontinuous conglomerate. On the eastern or dip slope of this part of the range the conglomerate has been pene-

trated by several wells and serves as an excellent aquifer. It is probable that this bed will furnish water to all homesteads under which it is present from Safford Peak to the north end of the range.

In the part of the range west of Tucson the most satisfactory wells appear to be on the larger cross faults. If the width of these fault zones is considered, it is not surprising that they serve as excellent watercourses and reservoirs. The eastward-trending valleys on the dip slope of the range are to a large extent localized on these faults, and in spite of the large run-off in the area much water from the summer and winter rains has an opportunity to soak into the faults.

Probably the most easily studied fault in the area is the one that follows San Juan wash eastward from the basin north of Cat Mountain. At the eastern mouth of the basin a dug well or shaft in the Cat Mountain rhyolite contained water to within 10 feet of the surface in April 1934. This well was apparently within a few feet of the fault. Farther east is a group of homesteads along the fault, and all the wells on these properties appeared to be satisfactory.

In addition to the water occurring along fault zones, water seems to be present to a lesser extent in the more-porous horizons in the Cat Mountain rhyolite, but, as would be expected, appears to be relatively scarce in the more-massive and dense volcanic formations. The water possibilities of the great thrust zone have not been adequately determined, but this appears to be a water horizon of some promise.

In connection with the development of water in the range, it appears significant that the shallower dug wells get more water than drilled wells. This is apparently due to opening more cracks and fissures with their larger cross-section and also to their larger storage capacity. It is believed that the ideal method of developing water would be to sink shafts on the faults to a depth where seepage of water became great and then cross-cut the fault. The cross-cut would open more fissures and add to the storage capacity of the well. In drilled wells of small flow, "shooting" the well should certainly be considered and may give a similar result.

A large proportion of the water for the city of Tucson is obtained from the underflow of the Santa Cruz River. The river channel is on the valley fill, and the elevation of the water in the underflow is influenced by the relation of the bedrock and the valley fill.

At San Xavier Mission, a basaltic ridge extends under the river, from the Black Hills to Martinez Hill, just east of the boundary of the map (Pl. 1). This ridge deflects the underflow of the Santa Cruz River to the surface, and at that point the Santa Cruz is a permanent stream.

Off the north end of the range, between the mountains proper and the small hill of metamorphosed Cretaceous Amole arkose north of the river,

there is apparently a ridge completely buried in the valley fill. Besides the alignment of the main ridge with the northern hill, the presence of the suballuvial ridge is shown by the height of the underflow. Smith (1910, p. 178) says that

> "the water table drops from a depth of 18 feet below the surface to a depth of 120 feet in a distance of less than a mile."

This might well be described as a suballuvial waterfall.

MINERALIZATION

Mineralization is widespread in the range. In the past the prospects of the area have undergone extensive development, but during the past few years there has been little activity. The total mineral production of the range is small. Owing to the inaccessibility of the workings the writer has not attempted to describe them in detail but instead can refer the reader to the descriptions by Jenkins and Wilson (1920, p. 15-25). The present study of the regional geology has brought out several features of the ore deposits which are of importance for further study.

The mineralization of the range occurred in two different periods—Laramide and Tertiary. In the vicinity of Amole Peak, around the Laramide intrusives, contact metamorphic deposits and veins are characterized by the presence of specular hematite and magnetite. The contact metamorphic deposits occur principally in the small Paleozoic limestone blocks along the great thrust fault and to a lesser extent in the calcareous horizons in the Cretaceous sediments. All the limestone blocks are too small to justify prospecting.

Fissure veins which usually carry specular hematite are common in the Cretaceous volcanic rocks north of Amole Peak. Most of these veins have a northeasterly strike. The Old Yuma mine is the best known and best developed of this system. It is noted as a locality for molybdate and vanadate minerals. The association of this mineralization with metamorphism of known Laramide age and the lack of veins of specular hematite and magnetite in the Tertiary volcanics indicate a Laramide age for this mineralization.

South of Amole Peak there are two centers of mineralization and contact metamorphism in the Cretaceous rocks around intrusives of unknown age. The contact metamorphism of the Cretaceous sediments in the hills southwest of Golden Gate Mountain and around Saginaw Hill is similar to that of Laramide age in the Amole Peak area. The intrusives are similar to those of the Laramide, and the mineralization is tentatively assigned to that period.

Cutting the Tertiary volcanic rocks are veins in which the high-temperature oxides have not been observed. Most of these are in the southern part of the range, and none of the prospect holes is open for study.

CONCLUSIONS

The foregoing account has outlined the structural and physiographic development of the range which is believed to reveal the most complete geologic history of the Arizona Basin Range mountains.

The age of the lavas capping the Basin Range block is almost certainly established as Tertiary, although the division of the Tertiary remains to be determined. Dating the lavas and the Cretaceous sediments establishes the age of the pre-Tertiary orogeny as Laramide, which differs from the late Jurassic age of the folding farther northwest in the Basin Range province.

Another feature in which the Tucson Mountains differ from those farther north in Utah, Nevada, and Oregon is in their great denudation since tilting and a lack of evidence of recent movement along the old structural lines. It may be suggested that the Basin Range ranges become progressively younger from south to north and are youngest in Oregon.

The work has shown that the two periods of diastrophism were very different, with Laramide folding and thrusting contrasting with Tertiary steep-angle tensional faulting. The later basaltic eruptions are believed to have issued from these steep-angle faults.

Pediments have been extensively developed during the long denudation of the range. The writer is inclined to the opinion that the slope of their bedrock surface is greater than generally believed and that their true gradient is best obtained from dissected pediments which have been stripped of their obscuring mantle of alluvium.

Evidence is presented which indicates a complicated history of erosion, burial, and stripping of the debris in which the range was buried. This involves a considerable complication of the physiographic history and needs further study.

Another point which requires further study is the age and relations of the Tertiary lake beds. These are believed to be pre-basalt in age and to rest on an old pediment. An alternative interpretation is that they are older and thrust on the Tertiary volcanic rocks. The problem can be solved by a determination of the age of the beds. If they prove to have come into position by thrusting, this will make a great change in future structural interpretations.

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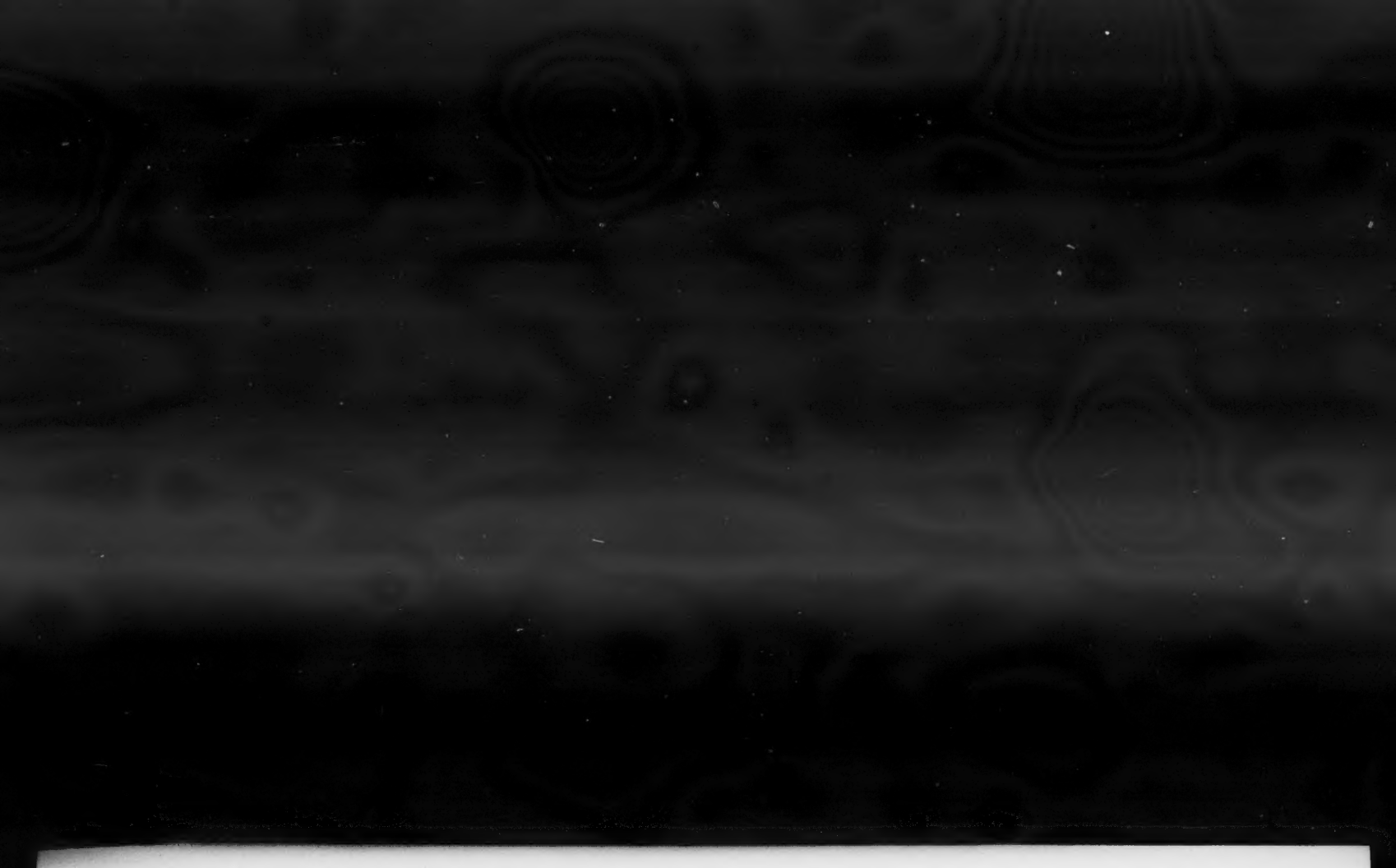

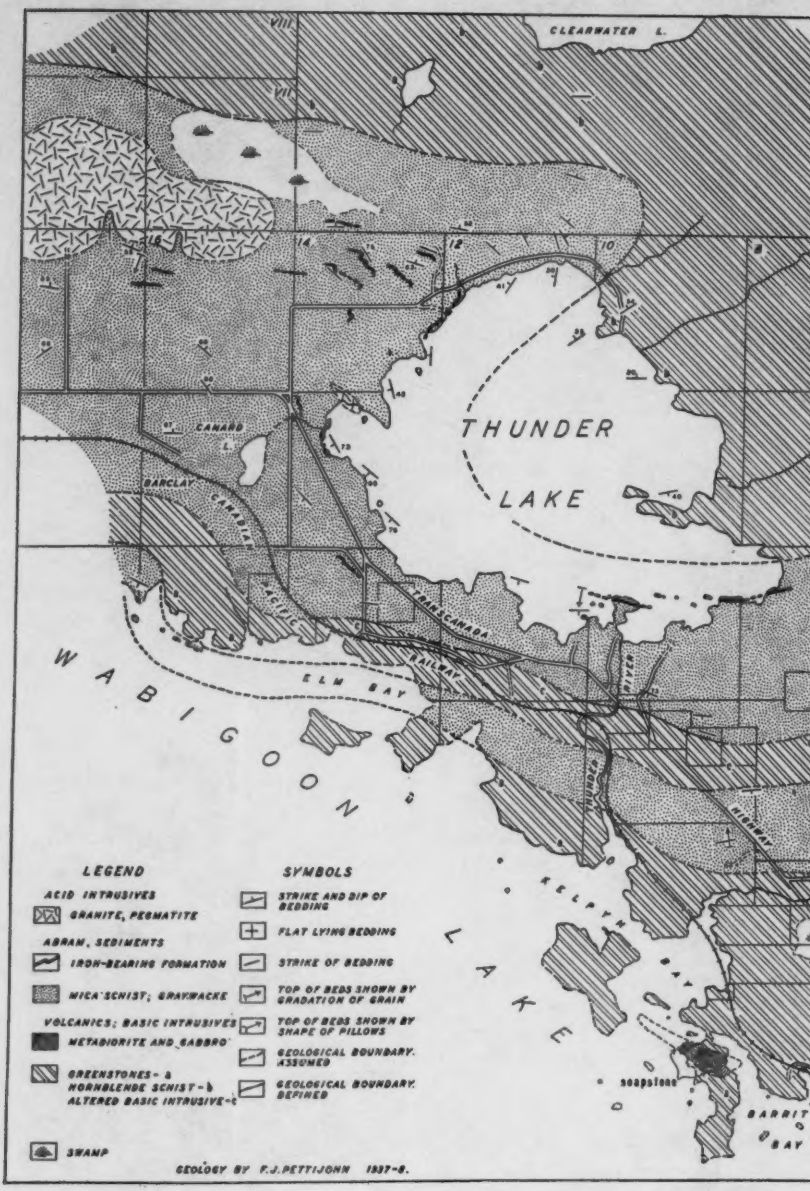

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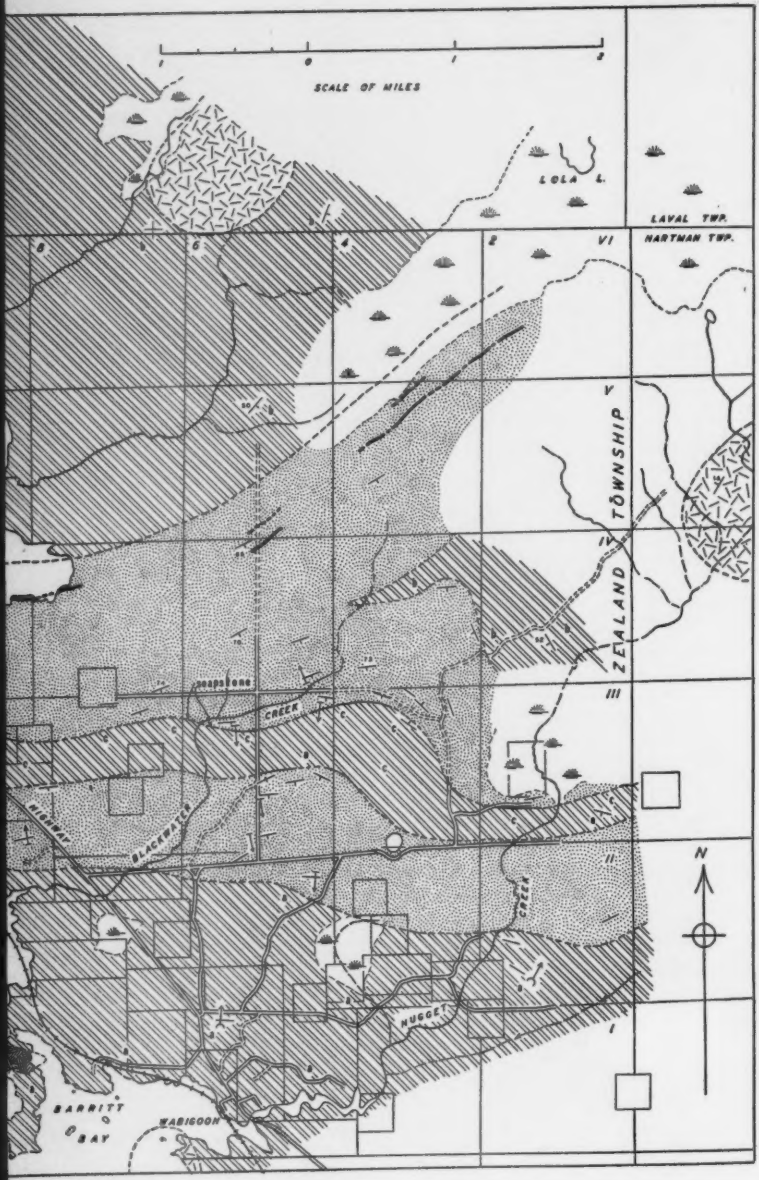
HUNDER LAKE AREA



HUNDER LAKE AREA

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"COUTCHICHING" OF THUNDER LAKE, ONTARIO

BY F. J. PETTIJOHN

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ABSTRACT

A mica schist series near Thunder Lake, District of Kenora, in northwestern Ontario, has been heretofore regarded as pre-Keewatin or Coutchiching. A field study of this series by the writer radically revises older mapping and proves the sediments to be younger than a Keewatin hornblende schist complex. The sediments, however, contain a minor greenstone flow series. A less metamorphosed greenstone series is also present, but faulting makes age relations of this series uncertain.

Basic intrusives are associated with greenstones and sediments. Batholithic granites and pegmatites cut all other rocks of this area.

INTRODUCTION

The relative age of the mica schists and associated greenstones in the Canadian shield is a perennial problem. In the Thunder Lake area (Fig. 1) of northwestern Ontario is a mica schist series designated Coutchiching on Map no. 266A (Kenora sheet) of the Geological Survey of Canada. These mica schists are clearly sedimentary. Their geologic

age, however, is not easy to determine. Coleman (1902) was apparently the first worker to assign them to the Coutchiching. McInnes (1902), on the Manitou Lake sheet of the Geological Survey of Canada, designated the formation as "highly altered Keewatin rocks, principally mica

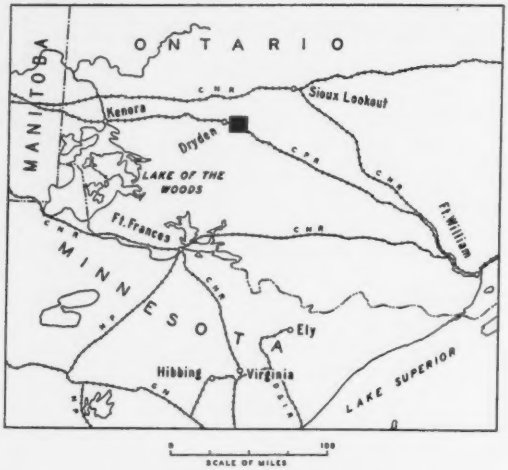

FIGURE 1.—*Map showing location of Thunder Lake area*
Black rectangle covers area of geologic map.

schists and fine gneisses." Parsons (1911) followed McInnes in assigning the schists to the Keewatin. Thomson (1917) described the sediments as "apparently a border phase of the Keewatin", though elsewhere he correctly interpreted them as due to "the metamorphism of a mixed argillaceous and arenaceous sediment." Since he described them under the heading "Keewatin Metamorphosed Sediments" it is clear that they were not considered an independent series. Wright (1924, 1926), in his report on the soapstone deposits of Wabigoon Lake and vicinity, concluded that "the general southward dip and the areal distribution of the sediments suggest that they underlie the volcanic series, and, therefore, are older." Apparently Wright's opinion was the basis of the assignment of the mica schists to the Coutchiching by Tanton (1933) on the Kenora map sheet of the Geological Survey of Canada. During a brief visit to Thunder Lake in 1935, the writer (Pettijohn, 1937) separately mapped a hitherto unknown belt of Keewatin hornblende schists. He concluded that these were older than the sediments because of their position in the center of

the Thunder Lake dome which is flanked by the sediments. The greenstones of Wabigoon Lake, however, were concluded to be younger than the sediments.

Most of the earlier geological work in Thunder Lake area, however, was of a reconnaissance nature and was hastily done. Such earlier work, as for example that of Coleman (1902), was done in conjunction with a survey of all the iron ranges of northwestern Ontario. Parsons' (1911) work was a part of a survey of all the active mines and prospects of the Lake of the Woods, Manitou, and Dryden gold areas. Wright's work in 1924 and 1926, the most recent original study, was done as a part of a study of the talc and soapstone deposits of Canada. Only the survey by Thomson was devoted specifically to the Thunder Lake and Wabigoon area. Even Thomson's work was more inclusive than the present one.

The study reported on here, since it is of a more intensive nature, has necessarily resulted in considerable revision of the geology of this area. Greater accessibility as a result of the settlement and road building has made possible easy discovery of many new outcrops. The new map (Pl. 1), therefore, differs materially from earlier published maps.

In addition to significant revision of the areal geology, the writer found further evidence on the relative ages of the paragneisses and the greenstones. The purpose of this paper is, therefore, to present the revised geologic map and to give the newer data on the Coutchiching problem of this area.

ACKNOWLEDGMENTS

Field studies were made by the author in 1937 and 1938. Mr. Arthur Lundahl rendered able assistance in field work both years. The study of the geology of the township was a part of a larger project supported by funds granted by the National Research Council and by a grant from the Salisbury Research Fund of the University of Chicago. To both sources of aid the writer gratefully expresses his thanks. Mr. E. L. Pidgeon, of Wabigoon, aided the progress of the work in many ways.

GENERAL GEOLOGY

GENERAL STATEMENT

The bedrock of the area falls into four major categories; (1) The greenstone or volcanic complex commonly called Keewatin, (2) the metasedimentary series or supposed Coutchiching, (3) a minor amount of basic material injected into and associated with the greenstone, and

(4) the various granitic intrusives. The formations are tabulated as follows:

Pleistocene
  Glacial drift
    Varved clay, outwash sands and gravel, sandy moraines
Early Pre-Cambrian
  Batholithic intrusives
    Granite, pegmatite
  Basic intrusives (in part contemporaneous with greenstone flows)
    Metadiorite, metadiabase, soapstone
  Abram series ("Coutchiching")
    Upper sediments: graywacke and phyllite
    Greenstone flows
    Lower sediments: mica schist, iron-bearing formation, graywacke
  Keewatin series (possibly in part post-Abram [in age])
    Greenstones, chlorite schists, hornblende schists

KEEWATIN (?)

The greenstones and their metamorphic derivatives crop out in four areas in the vicinity of Thunder Lake (Pl. 1). Since the southernmost area underlies most of Wabigoon Lake and the area about the village of Wabigoon, it may be termed the Wabigoon belt. The contact with the sedimentary mica schists skirts the north shore of the lake and runs east inland from the lake at about the mouth of Thunder River. This southern area is the most typically "greenstone" in type. Massive greenstone predominates, though locally between flows or along shear zones the rocks are schistose. Ellipsoidal or pillow structures are abundant. The Wabigoon greenstones are cut by somewhat altered gabbros and diorites in several places, notably in the village of Wabigoon and on the west shore of Barritt Bay. Alteration of still more basic rocks has produced the soapstone deposits of Barritt Bay. The Wabigoon greenstones are also cut by quartz porphyry dikes near the village of Wabigoon and in the southeast corner of Zealand Township, especially in Concession II.

A second and narrower belt of greenstone lies a short distance north of the greenstones just described and is separated from them by a narrow belt of sediments. This belt may be referred to as the Thunder River belt since it is well exposed at the Thunder River crossing of the Canadian Pacific Railway and in road cuts near the highway bridge. This greenstone band crops out on the north shore of Elm Bay, follows the railroad east to Thunder River, and extends eastward to the township limits in Concession III. The band is narrow, scarcely a quarter of a mile wide on the average, and is rarely, if anywhere, more than half a mile wide. Yet it forms conspicuous high ridges in lot 1 and lots 4 to 9 of Concession III. The rocks of this belt are of two kinds, namely, schis-

tose ellipsoidal greenstone and sheared and altered intrusives of diorite and more basic rocks. Were it not for some outcrops of pillow lava the whole belt might be mistaken for a great sill. The flow greenstones are nearly everywhere confined to the south half of the band. These pillow lavas are darker and more hornblendic than those of the Wabigoon belt.

A third area of volcanic schists, separated from the two southern areas described by the main belt of paragneisses, extends from the northeast corner of Zealand Township in Concession VI to Thunder Lake, where it swings north and west into Concessions VII and VIII of Zealand addition. It appears to end in lot 19 of Concession VIII. This band, not heretofore shown on any geologic map, differs from the others in that it is more highly metamorphosed. The great granite batholith with which it is in contact has materially altered the greenstones. The rocks as a result are much blacker, more coarsely crystalline, markedly hornblendic, and more strongly banded. They are largely hornblende schists rather than greenstones; nevertheless, their original nature is never in doubt. Vestiges of the pillow structure remain. Numerous large and small sharp-walled dikes of granite cut the hornblende schists, as do many pegmatites. The hornblende schists of this belt, which may be termed the Clearwater belt (from the high outcrops around Clearwater Lake), are well exposed at several points along the northeast shore of Thunder Lake. The rocks here are remarkable for their low-dipping schistosity; in places dips are as low as 20 degrees.

Garnetiferous hornblende schists also crop out in lots 1 and 2 of the fourth concession of Zealand Township. This Keewatin area is apparently separate from the other areas and is rather highly metamorphosed. The rocks contain abundant large irregular garnets.

ABRAM SEDIMENTARY SCHISTS (COUTCHICHING ?)

The supposed Coutchiching series of Thunder Lake consists of mica schists and paragneisses of sedimentary origin. Two belts of outcrops, oriented roughly east-west, cross the township. The lesser and southernmost one separates the Wabigoon greenstones from those of the Thunder River belt. This belt of sediments is narrow, averaging half a mile in width, and is generally poorly exposed. The best outcrops may be seen along the road that follows the line between lots 5 and 6, Concession III, and near the Transcanada highway about 1½ miles west of this place. The sediments of this belt, graywackes and phyllites, alternate with each other in fairly regular fashion. In places the sediments are much sheared and are converted to glossy gray to greenish-gray schists. Bedding occurs in all outcrops, though it is obscure in many. Crumpling of the bedding

is common, as are numerous small and irregular knots and veinlets of bluish-gray quartz.

The main body of sediments, the Thunder Lake belt, is more varied in composition. Quartz-biotite schist is most common, though many of the altered sediments are paragneisses. In spite of metamorphism, however, the original differences in grain size and mineral composition between one bed and another are faithfully retained. Original sands and muds, alternating in thin beds, were converted first to graywackes and slates and then with increasing metamorphism to quartz-biotite schist and paragneiss. Staurolite is present in some of the mica schist and is an index of a higher grade of metamorphism than is common in this region.

In the synclinal belt just north of Thunder Lake near the north line of lot 11 the sediments are singularly well exposed (Pl. 2, fig. 1). They are dark, thin-bedded, and fine-grained. Original bedding features, even paper-thin laminations, are well preserved. The metamorphism is of a lower grade than is common for this region, for elsewhere many schists are coarsely micaceous and locally carry staurolite and garnet. The lesser metamorphism of the fine-grained dark rocks is perhaps attributable to the carbon in the sediments. Eskola (1932) has, for example, noted the "protective effect" of graphite in certain schists in Finland. The sedimentary facies represented in the outcrops suggest strongly the Phantom Island formation of the East Bay (Minnitaki Lake) area where the rocks are also thin-bedded, varve-like, carbonaceous, and fine-grained.

In the sedimentary schists just described are nodules, a few inches in diameter, which probably were once calcareous concretions. They are now, however, completely silicated and consist of a fine-grained mixture of quartz and labradorite in which are numerous blades of hornblende. A little garnet and apatite are also present. No trace remains of the original carbonate. Some beds, a few inches thick, composed of a mineral assemblage similar to that of the concretions are interbedded with the normal mica schists. These are interpreted as calcareous shales which have likewise been completely silicated. Both the nodules and the limy beds are more resistant to weathering and stand up in slight relief on the outcrop (Pl. 2, fig. 1). The concretions tally closely in appearance and composition with those described and analyzed by Eskola (1932).

Interbedded with the paragneisses is an iron-bearing formation marked by layers a quarter of an inch to half an inch thick of magnetite interbedded with ordinary mica schist and garnet schist and with layers of sugary quartz. The iron-bearing formation is persistent and may be traced southwest from lot 3, Concession V, of Zealand Township to the southeast bay of Thunder Lake. Here it outcrops on the mainland and

FIGURE 1. VERTICAL BEDS IN SYNCLINE NORTH OF THUNDER LAKE
Small ridges on the outcrop are silicated calcareous layers.

FIGURE 2. SOUTH-DIPPING BEDS
On small island near south shore of Thunder Lake.

FIGURE 1. PEGMATITE CUTTING TOURMALINIZED SEDIMENTARY SCHISTS
Bedding is accentuated and darkened by tourmaline replacement. Area near knife is unreplaced schist. Note corroded contact of pegmatite with schist—sharp in some places, vague in others. Signal Hill pegmatite.

FIGURE 2. PTYGMATIC FOLDS
Pegmatitic granite cutting mica schist. Signal Hill pegmatite.

FIGURE 3. SILL-LIKE GRANITE SHEETS
Intruding nearly horizontal Keewatin hornblende schists. Roof (?) of Hartman granite.

PEGMATITES AND GRANITE

on the several small islets in this bay. Thence the formation swings northwest beneath the waters of the lake to reappear on the far west and northwest shores. Coleman (1902) early noted and described the conspicuous outcrops of a much-contorted iron-bearing formation in the north half of lots 13 and 14 of Concession VI. From that point it was traced west by the writer to the north half of lot 17 of the same concession. The formation was again seen several miles farther west in cuts along the Transcanada highway less than a mile east of the town limits of Dryden.

The formation contains enough magnetite to make the compass unreliable in the immediate vicinity of the deposit. At no place, however, was the formation rich enough to warrant prospecting for iron ore. Most characteristic of the outcrops are the contorted layers. No other formation shows drag folding so well or so commonly as does this rock. Noteworthy also in the Zealand area is the association of the iron-bearing rocks with layers of schist studded with small garnets. The concentration of garnets along certain layers shows clearly that original composition and grade of metamorphism are responsible for their formation.

The sediments of the Thunder Lake area are known to extend northeastward for many miles and are, in fact, a direct continuation of the Abram series of Minnitaki Lake described elsewhere by Hurst (1933) and by Pettijohn (1934). The principal differences between the sediments in the vicinity of Thunder Lake and those of Minnitaki Lake result largely from different degrees of metamorphism. A marked and progressive increase in metamorphic grade occurs between the Red Pine Bay area and Zealand Township. This greater metamorphism is correlated with an increased number of granite dikes and stocks in the latter area as well as with the proximity of the batholithic granites of the northern part of the township. Accompanying the increase in metamorphic grade are a disappearance of the axial-plane cleavage of the Minnitaki area and the appearance of a simple bedding-plane schistosity.

The paragneisses of Thunder Lake are known to continue westward to Eagle River and to Eagle Lake itself where they apparently end. The associated iron formation is reported at Eagle River by Parsons (1911).

BASIC INTRUSIVES

The basic intrusives superficially resemble the flow greenstones. They are dark rocks, many altered to a greenstone, composed of a plagioclase feldspar commonly saussuritized, an amphibole probably derived from a pyroxene, and much chlorite. In general they are coarser-grained and more massive than the normal greenstones and completely devoid

of pillow structures or any other structures suggestive of a flow origin. Some are very coarse. Despite these differences it is difficult to distinguish the intrusives from flows, especially when both rocks are strongly recrystallized and foliated. It was not possible to separately map the basic flows and intrusives in the Thunder Lake area because of scarcity of outcrops and lack of time. Therefore most of the basic intrusives are mapped with the volcanic complex despite the younger age of the intrusives—an age which is in part, at least, post-sedimentary.[1]

The original nature of these altered intrusives has not been worked out in the Thunder Lake area. In the Minnitaki Lake area and in the Abram and Little Vermilion lakes area, however, similar intrusives prove to be granophyric diabases and diorites.

Locally dikes of ultrabasic rock intrude the metadiabases and metadiorites. The ultrabasic rocks have been completely altered to soapstone. Such soapstone "dikes" are exposed on the irregular peninsula between Barritt Bay and Wabigoon Lake and also along the north edge of the Thunder River greenstone belt in lot 6, Concession III.

The basic intrusives are post-Abram. In the north part of lot 4, Concession III, an unfaulted and unsheared contact between the intrusives and the sediments is exposed. Within a few feet of the contact, gradations in grain in the sediments indicate that the top of the beds are toward the igneous rock. No conglomerate nor any other evidence exists to indicate a superposition of the sediments on the intrusive. An intrusive relation is substantiated in the Red Pine Bay area of Minnitaki Lake where sills of similar rocks clearly invade the sediments. Contacts between the basic intrusives and the granites of this area are not known. The basic rocks are, however, much sheared and were apparently involved in the diastrophism of the region. Since the granites are in the main undeformed, they were intruded after the diastrophism and are therefore younger than the basic intrusives.

BATHOLITHIC INTRUSIVES

Granitic rocks invade both the sediments and volcanics in many places. Several stocks have been discovered and partially mapped in addition to the great batholithic complex which lies just north of the map area.

One small stock lies in lots 17 and 18 of Concession VII. This stock was described by Coleman (1902) who thought it was a part of the batholith of the Gullwing Lake area north of Zealand Township. The existence of this stock seems to have been entirely overlooked by later

[1] The only basic intrusive separately mapped is that on the irregular peninsula which separates Barritt Bay from Wabigoon Lake. This mass was separately mapped and described by Wright (1926).

workers, though Coleman's description is very accurate and complete. The intrusive is largely of pegmatite but also contains some granite. This stock may be called the Signal Hill pegmatite from a high knob of the same name on which this rock is well exposed. The pegmatite is chiefly quartz, microcline, and muscovite. Much of the quartz and feldspar is graphically intergrown. Black tourmaline is a common constituent and is locally very prominent. Small red garnets are accessory at a few outcrops. The mica has a plume-like structure in the coarsest facies of the rock. Within the paragneisses, into which the stock is intruded, and near the borders of the stock are many small and large dikes of pegmatite, mineralogically similar to the pegmatite of the stock itself. These dikes are of two kinds. The first group are parallel to the bedding, and many have a "colon" shape in outcrop plan. They have irregular walls and are curiously constricted at certain points. The second kind of pegmatite runs across the bedding. These dikes are in some places markedly "sigmoidal" or show what Sederholm calls "ptygmatic folding" (Pl. 3, fig. 2). The folds range in amplitude from a few inches for dikes about an inch wide to several feet or even yards for dikes a foot thick. The very large dikes, some tens of feet across, do not seem to be folded. Very probably both the colon shape and the sigmoidal folds of the pegmatites were formed by deformation of the country rock after the injection of the dike material and before its complete consolidation. Pressure applied normal to the bedding (and schistosity) resulted in shortening by folding of any transverse dikes. There was also a corresponding development of tension parallel to the bedding resulting from the flow of matter accompanying the compression. The dikes parallel to the bedding, therefore, would normally have been pulled apart and would develop a series of transverse fractures—in effect would form ladder veins. But, inasmuch as the dike material was not solid, it failed to sever sharply and instead thinned in the same manner in which soft material thins on stretching. The colon-like constrictions observed were such points of thinning. That the country rock was in a quasi-liquid condition also is shown from the manner in which it is "infolded" at these same constrictions.

Further evidence of deformation of these dikes lies in the broken tourmaline crystals. Tourmaline crystals, half an inch in diameter and 4 inches or more in length are severed in three or more places. The severed fragments are more or less separated but are still aligned. The breaks are healed with quartz.

Tourmaline occurs not only in the pegmatites but in the mica schists into which the dikes are intruded. The tourmaline in the schists is fine-grained and acicular and is very abundant in a zone a few inches wide

adjacent to the dikes and in inclusions within the dikes. The schists are locally almost wholly replaced by tourmaline. The original bedding of the schists is preserved and even accentuated (Pl. 3, fig. 1).

In the northwest corner of Hartman Township, lot 11, Concession V, are some light-gray granite outcrops not heretofore mapped. An acid plagioclase, quartz, and a little biotite are the minerals visible to the naked eye. Faint lamination and some granulation of the feldspars indicate deformation after intrusion. One remarkable feature of the outcrops is the large sheets of hornblende schists, derived from the greenstones, which are interleaved with the granite. One is at a loss to know whether to describe the granite as a series of sill-like intrusions into the Keewatin or to describe the Keewatin as a series of nearly flat, large inclusions (Pl. 3, fig. 3). The dip is slight, in many places nearly horizontal. Very probably, then, the granite now exposed was near the roof of a dome-like stock. The full extent of this granite is not known, but it is believed, however, to be an isolated stock of no great size.

Just north of lot 6 of Concession VI of Zealand Township, about 2 miles from Thunder Lake, is another granite body. This is of reddish color, medium-grained, and composed of feldspar, both acid plagioclase and reddish orthoclase, much quartz, and very little biotite. The granite is largely drift covered so that its relation to other rocks is not clear. It is located near the center of the Thunder Lake dome and is probably in some manner related to that structure. The full extent of this stock is unknown but is believed to be small.

A small stock of pink granite crops out on the southeast shore of Barritt Bay in Concession I of Zealand Township and extends into the northwest corner of Southworth Township. According to Thomson (1917), the granite is a hornblende-biotite granite.

A granite batholith crops out in the north part of Zealand addition and in the territory east of there and north of the map area. This granite is of coarse grain and light color and is generally unfoliated. A specimen from lot 5, Concession II, of nearby Wainwright Township, is described by Thomson (1917) as a coarse-grained aggregate of quartz, orthoclase, plagioclase, biotite, musicovite, and epidote with minor amounts of magnetite, apatite, and chlorite.

STRUCTURE AND AGE RELATIONS

All the rocks of the Thunder Lake area have been deformed by folding as well as altered by recrystallization. The recrystallization has produced a schistosity universally parallel to the bedding. The symbols used on the map, therefore, to denote strike and dip of "bedding" also record the strike and dip of the schistosity. The axial-plane cleavage of folded

rocks so commonly observed in the Minnitaki region in which these same beds are exposed apparently disappears with recrystallization (insofar as the naked eye is concerned) and is replaced by a schistosity due to recrystallization and orientation of mica and hornblende crystals along bedding planes.

The most conspicuous structural feature of the area is the Thunder Lake anticline which is at once made apparent by a study of the dips of the hornblende schists of the northeast and east shores of Thunder Lake. On the prominent peninsula of the south halves of lots 9 and 10, Concession V, the hornblende schists dip south at an angle of 40 degrees, whereas schists of the same kind a mile farther north along the east shore dip 20-35 degrees to the north. The anticline thus proven plunges sharply westward so that only sedimentary mica schists and iron formation crop out elsewhere around the lake. This westward pitch is also expressed in the swing in strike of the sediments from nearly east-west through northwest, and north, to northeast as one follows the shore from the outlet of the lake around the south, west, and northwest sides of the lake.

Farther east, just north of lots 6 and 7, Concession VI, a granite stock is located on the axis of the fold and is apparently related in some manner to the structure.[2]

The Hartman stock, already described, is probably along the axis of another anticline. Less careful survey and fewer outcrops, however, make it impossible to describe its full extent and nature. Near its northwest margin hornblende schists dipped 14 degrees away from the granite.

No very clear explanation has been advanced to account for the concordant relations of the granite and the flanking schists on dome-like folds which contrast sharply with the usual discordant relations observed between granite and schist. Nor is there any obvious reason why sediments elsewhere slightly altered and folded into tight isoclines should here be so much more highly metamorphosed and yet form such gentle open folds. A few writers have commented on similar structures. Billings (1937, p. 522), for example, observed a similar fold, the Owls Head dome, of larger size, in the Littleton-Moosilauke area in New Hampshire. He attributed the structure to injection of the granite before the main folding. The dome served then as a buttress controlling the younger

[2] Since greenstone is reported at Rafter Lake several miles farther northeast, it is probable that the granite is an isolated stock rather than an extension of the batholith of Gullwing Lake. It is probable, too, that hornblende schist will be found to dip northeastward as well as southeast, south, west, and northwest. The anticline is therefore probably an elongate dome analogous to the Rice Bay dome of the Rainy Lake area described by Lawson, except that in that area the mica schists flanked the granite core and were in turn flanked by the Keewatin.

structures. Lawson (1913, p. 56) regarded the granite in the Rice Bay dome as injected, perhaps as a flat sill, into flat-lying Coutchiching schists. The same forces responsible for the development of the dome in the schists involved the granite and produced the structure seen.

The sedimentary schists which lie on the northwest flank of the Thunder Lake anticline are bent sharply back on themselves and show also an abrupt change in strike from northeast to northwest and west. This synclinal structure has its axial plane along the north line of Concession VI. The southern limb is also the north side of the Thunder Lake structure and accordingly dips north. The northern limb is overturned and the beds, therefore, also dip north beneath the Keewatin of the Clearwater belt. The actual contact between the mica schists and the hornblende schists was seen only in lot 18, Concession VIII. The contact is sharp. No conglomerate is present, nor is there any other indication of the relative age of the two formations. One must rely, therefore, wholly on the map relations described to determine the order of succession. Most remarkable is the steep east dip of the beds on the easternmost point of the "nose" of the fold. The syncline is shown by this eastward pitch to be inverted. Such inversion is not unique since a similar inversion is described by Billings (1937) in his paper on the Littleton-Moosilauke area. The mechanics of such inversion are not known.

The syncline becomes a simple "drag" or bend in the strike of the beds farther southwest. The sharp bend in strike near Barclay is an enfeebled expression of this fold.

The structure of the rocks south and southeast of Thunder Lake is not clear. The two belts of sediments might be interpreted as synclines downfolded into a basement of Keewatin rocks, but field data are not consistent with this view. Effort was made to determine the position of top and bottom of the near-vertical beds. The gradation of grain size commonly shown between the bottom and top of a single bed was utilized, as was the shape and manner of packing of the pillow structures in the lava flows. Data from these sources are shown on the map (Pl. 1) by appropriate symbols. Inspection of the meager data yields no simple interpretation. The best evidence gleaned from a study of outcrops along the road between lots 5 and 6, Concession II, suggests an interbedded position for the narrow Thunder River belt of greenstone. The north contact of this greenstone and the sediments may be seen in lot 4. Here, however, the greenstone is intrusive, and the relation to the true flow greenstone of this belt is obscured. These coarse-grained basic intrusives, nearly universally associated with Keewatin type rocks, are regarded by most writers as a facies of the Keewatin volcanism. If the

greenstone flows were submarine, the rising magma would tend to spread under the cover of already consolidated lava. Such behavior is a special case of the common practice of lava to flow through tunnels of its own consolidated substance. Such auto-intrusions, though commonly recognized, are difficult to map separately from the associated flows. If the basic materials of the Thunder River belt are so interpreted it is clear that they were intruded between a floor of sediments and a cover of flow rocks. That the intrusives were injected before deformation is proven by shearing and foliation which the rock exhibits in common with the pillow lavas.

The sedimentary beds north and south of the greenstone belt apparently are oriented in the same direction, namely with their tops facing south. Pillow structures in lot 4 suggest a similar orientation for the greenstones. The sediments of the main belt and those of the narrower southern belt would therefore be interpreted as of different ages separated by a thin flow series. The absence of iron formation in the southern band of sediments in contrast to the abundance of such formation in the northern band—a situation not accounted for by the downfolded synclinal interpretation—is thereby explained.

The apparent bifurcation of the main band by the hornblende schists of lots 3 and 4 of Concession IV is difficult to explain. Further work will be required to account for this complexity.

The Wabigoon greenstone belt has well-formed pillows in some places. Wherever noted, these vertical beds seem to show the tops of the flows to be north-facing. This position is apparently inconsistent with the generally south-facing position of the also near-vertical sedimentary schists. Very probably, therefore, the contact between the sediments and greenstones is the plane of a major shear, and because of the fault relations no certain conclusions can be drawn concerning the relative age of the Wabigoon greenstones and the Abram sedimentary band. The faulted nature of this contact is confirmed by a study of the rocks near the contact. Just west of the diagonal road in the north half of lot 5, Concession II, for example, is a greenstone outcrop—the one nearest the projected position of the contact—which contains greatly elongated and much-sheared pillows. Farther west along the north shore of Wabigoon Lake, the rocks are highly sheared and are very schistose. The schistosity, however, is induced by shearing rather than by recrystallization. Simple shearing, of the type postulated along the greenstone-sediment contact described, produces a thin "papery" schist—fine-grained and sericitic in the sediments and acid lavas and equally fine-grained and chloritic in the greenstones. Carbonate solutions readily invade these

nacreous and glossy green schists and replace these schists to a substantial extent.

SUMMARY

The relations of the sedimentary schists to the volcanic schists have been shown to be complex. Hornblende schists of volcanic origin are definitely older than the supposed Coutchiching mica schists. The occurrence of the volcanic schists in the axial region of the Thunder Lake anticline and their low dips beneath the flanking sedimentary schists are proof of the conclusion given. A relatively thin but persistent greenstone series divides the sedimentary series into an upper and a lower division. The greenstones crop out as a narrow belt separating the two sedimentary schist series. The interbedded position is shown by an unfaulted contact with the lower sediments where graded bedding proves the top of the sediments to be south-facing toward the greenstones, and by a similar south-facing orientation of the tops of the greenstones and the tops of the superjacent sediments. The greenstone belt cannot be interpreted either as an anticline or syncline because of the top-bottom relations described, and because such interpretation would require a repetition in the southern sediments of the iron-bearing belt known to occur in the sediments north of the greenstones in question. Such repetition does not exist.

The markedly schistose zone near the contact and the fact that the tops of the flows face north while the tops of the sediments face south are evidence of fault relations of the greenstones of the Wabigoon area with the upper sediments.

The facts just reviewed do not permit an assignment of the mica schists to the Coutchiching if the latter be defined, as it was by Lawson, as a pre-Keewatin sedimentary series. Reconnaissance mapping demonstrates a direct connection of these Coutchiching-like schists with the sediments of Knife Lake type in the Minnitaki Lake area. The difference in grade of metamorphism between the two areas is directly related to the proximity of acid plutonic intrusives.

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THE PEWABIC MINE

BY HARRISON SCHMITT

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ABSTRACT

The Pewabic mine is in the southeastern part of the zone of pyrometasomatism which surrounds the intrusive quartz monzonite at Hanover, New Mexico. Extensive development has proved that the south lobe of the intrusive is floored and that the sedimentary rocks, chiefly limestones, shales, and intercalated quartz diorite sills,

were thrust aside laterally by the magma. The overthrust, a minimum of 500 feet, was expressed by close folding accompanied by rock flowage of the bedded rocks in the upper plate of a flat thrust fault.

The less intense effects of the exomorphism resulted in marbleization of the limestone and hornstonization of the shales; the stronger effects were epidotization of the aluminous rocks and andraditization of the limestones. The lower plate was mildly metamorphosed. The sphalerite ore bodies, essentially uncontaminated by lead and copper, are localized by the intersection of the thrust fault with nearly vertical post-silicate northeast fault zones resulting in flatly pitching chimneys and ore blankets.

The facts indicate that the rocks were metamorphosed without volume change and that large amounts of silicon, iron, and zinc were brought in. Much of the silica can be accounted for by epidotization and chloritization of the aluminous rocks, and a question discussed is whether the silica, largely older than the sphalerite, may not have been derived from alterations by gaseous volcanic emanations.

INTRODUCTION AND ACKNOWLEDGMENTS

This paper describes the sedimentary and igneous rocks, geologic structure and contact pyrometasomatism of the Pewabic mine area, Hanover, New Mexico, and discusses some implied generalizations. The area, part of the Central Mining District, is the southeast section of the zone of metamorphism which surrounds the Hanover monzonite intrusive (Fig. 1).

The Pewabic mine exploits a deposit of sphalerite associated with pyrometasomatic silicates. The ore which is essentially lead-free, thereby commanding a premium for the manufacture of high-grade spelter, zinc alloys, and zinc oxide, is low in copper and other metals. (*See* Table 5.) The deposit lies in an overthrust plate of closely folded and overturned sedimentary rocks and in an intercalated sill of diorite porphyry (Pl. 2, c). The thrust fault at the base of the upper plate lies within the mine workings and marks approximately the lower limit of the silicate metamorphism (Pl. 2, section AA′).

The chemical analyses of the rocks and minerals, the spectrographic analyses, and some of the thin sections were made possible by a grant from the Penrose Bequest of the Geological Society of America. Large-scale maps of the surface and underground geology were prepared by the writer while employed, during the past 5 years, by the Peru Mining Company which owns the Pewabic mine. This work was a continuation and extension of studies of the Central Mining District carried on intermittently by the writer since 1927. Thanks are due the Peru Mining Company for permission to publish data collected from its property.

SEDIMENTARY ROCKS

The sedimentary rocks exposed at or near the Pewabic mine include about 1100 feet of the Central Mining District succession which totals

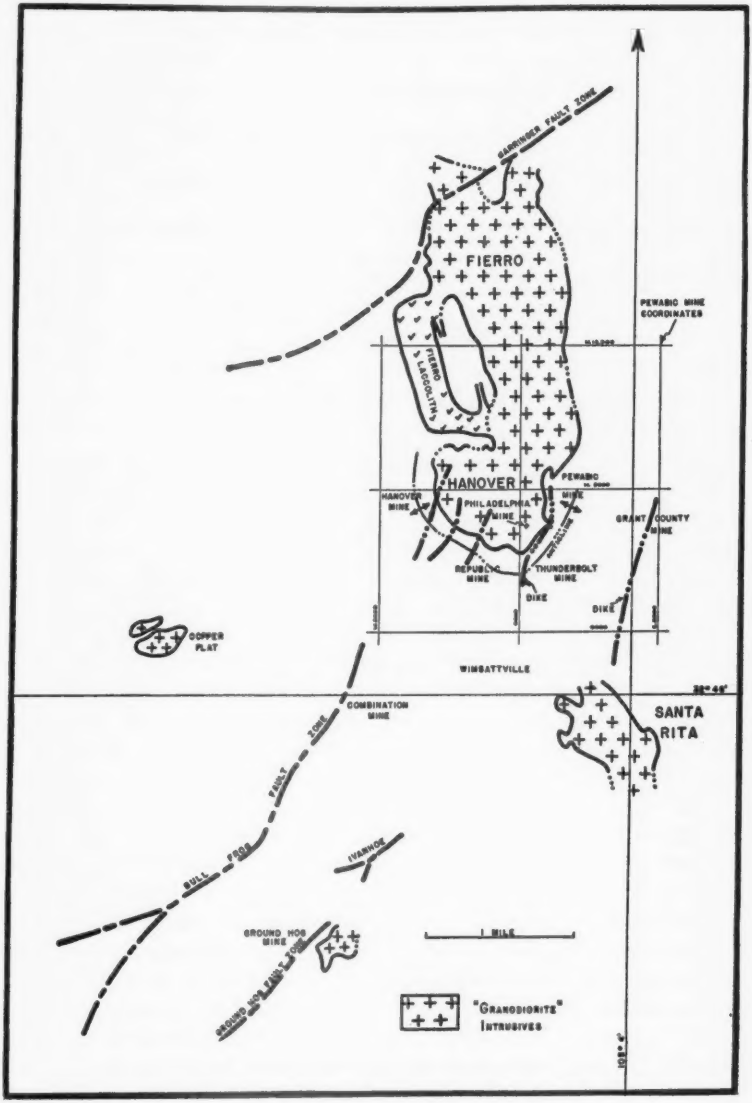


FIGURE 1.—*Key map of the Central Mining District*

5000 feet. The local rock column with current local nomenclature, including the Hanover diorite porphyry sill, is as follows:

PEWABIC MINE SECTION*

Age	Regional Name	Local Name	Thickness (feet)
Lower Pennsylvanian	Lower Magdalena	Upper Blue limestone	350
Laramide (?)		Hanover sill	65–85
Lower Pennsylvanian	Lower Magdalena	Middle Blue limestone	85
" "	" "	Parting shale	18
Lower Mississippian	Lake Valley limestone	Hanover limestone	110
" "	" " "	Lower Blue limestone	190
Upper Devonian	(fossil horizon)	Transition beds	100
" "	Percha shale	Percha shale	200
Silurian (Niagaran ?)	Fusselman limestone	Fusselman limestone	50

* For further detail and a revised nomenclature, *see* Paige (1916), Spencer and Paige (1935), Lasky (1936), and Schmitt (1935b), p. 188.

The Fusselman limestone is Silurian, apparently Niagaran, in age, for it contains the corals *Halysites catenulatus, Favosites niagarensis* (?), and, according to Paige, *Pentamerus* sp. It is a fine-grained, massive-bedded magnesian limestone separated from the Percha shale by an unconformity, apparently a disconformity, for no angular discordance has been noted in the district. The lower part of the Percha shale is black and fissile, and regional studies suggest its Upper Devonian age. The upper 100 feet, usually mapped with the lower Percha shale, is marked by thin fossiliferous limestone beds and nodules which increase in abundance and thickness upward until they form the massive Lower Blue limestone. According to Kindle (Paige, 1916, p. 5), the fossils in this zone are of Upper Devonian age.

The Lower Blue limestone is medium bedded, shaly, and non-magnesian. (*See* Table 9.) The beds are gull gray, the depth of color being proportional to the content of shaly and carbonaceous matter. Thin shale beds occur in the upper three-quarters, and a minor amount of gray chert is conspicuous near the top. The lower quarter contains little impurity, is recrystallized to white marble near intrusives, and has been replaced by small low-grade ore bodies in neighboring mines.

The Hanover limestone, where unmetamorphosed, is massive, coarse, white to grayish-white, non-magnesian, and composed largely of crinoid stems. The only important impurity is about 5 per cent white chert which occurs in lenses. When the chert is excluded the limestone is nearly 100 per cent calcium carbonate. (*See* Table 9.) The Hanover limestone is more susceptible to replacement than the other lithologic

units of comparable thickness in the district and contains a large proportion of the zinc ore. The lower contact is marked by a change downward from a white calciphyr to fine gull gray [1] limestone and from white to gray chert. No shaly impurity, either disseminated or in layers, has been observed in this limestone.

The Parting shale rests disconformably on the Hanover limestone, the contact marked locally by a thin conglomerate which is in places mainly white chert derived from the underlying Hanover limestone. Where unmetamorphosed the shale contains abundant plant remains, including *Sigillaria* sp. The microscope reveals a greater proportion of fine, rounded sand grains than macroscopic inspection suggests. Some layers should be classified as fine-grained sandstone which, near the main intrusive, is metamorphosed to quartzite. (*See* Table 7, analysis P-27.) The Parting shale nowhere contains important quantities of zinc ore.

The Middle Blue limestone is purer than the Lower Blue limestone but nevertheless closely resembles it. (*See* Table 9.) The lower half is fossiliferous and approaches the Hanover limestone in purity and appearance. The upper half is gull gray because of shaly impurity and contains several 2- to 4-foot shale layers. The Middle Blue limestone contains a large proportion of the ore in the Pewabic mine but includes mostly low-grade ore bodies in the neighboring Empire Zinc Company's property.

The Hanover sill, like most of the sills of the district, is intruded into a shale bed. An impure coal layer about a foot thick occurs at the bottom of this shale in the Pewabic mine and at several other places in the district, and nowhere more than 15 feet below the bottom contact of the sill. The upper part of the shale is in many places at the upper contact of the sill, and the overlying limestone is in places marked by a 5- to 10-foot zone containing an exceptionally high concentration of black chert which encloses abundant *Fusulina* sp.

The Upper Blue limestone is gull gray because of minor shaly impurity. Shale occupies about 10 per cent of the interval. The limestone is massively bedded and nonmagnesian. (*See* Table 9.) About 60 feet above the 10-foot black chert already mentioned in a zone which characteristically contains *Chaetetes* sp.

Most of the sedimentary rocks of the Pewabic mine were, before metamorphism massive-bedded, slightly shaly, gull gray limestones with intercalated shale beds comprising about 10 per cent of the section. Such limestone would normally be classed as structurally competent. As will be noted later, however, it was incompetent under the conditions which produced the local structure.

[1] Ridgway scale.

IGNEOUS ROCKS

The igneous rocks of the Pewabic mine area can be classified into three age groups: (1) The early quartz diorite porphyry sills, laccoliths, and sheets, (2) the later granodiorite [2] and granodiorite porphyry which almost everywhere cross-cuts the structure, and (3) the late vitrophyric dacite porphyry post-ore dikes. This sequence of igneous rocks is typical for the Central Mining District. The late vitrophyr cross-cuts at all known exposures. Sills, sheets and laccoliths of quartz diorite porphyry, and dikes and chonoliths of granodiorite and granodiorite porphyry intrude the Colorado shale but are truncated by a prominent widespread angular unconformity probably representing early Tertiary (Eocene ?) erosion. Not more than 4 miles from the Pewabic mine area this unconformity is overlain by beds of gravel, sand, and tuff, many of which appear to be lacustrine. They contain numerous fragments of the quartz diorite and granodiorite and contact metamorphic rocks and ores.

TABLE 1.—*Chemical analyses of igneous rocks of the Pewabic mine area*

(A. Willman, analyst; direction, F. F. Grout)

	P–20*	P–21**	P–19†	P–22††	Average
SiO_2	58.89	64.15	65.36	64.53	63.23
Al_2O_3	16.55	16.45	16.29	16.11	16.35
Fe_2O_3	.55	1.13	1.94	1.50	1.28
FeO	4.56	2.12	1.88	1.56	2.53
MgO	2.72	2.05	1.76	1.39	1.98
CaO	7.60	7.09	4.05	2.81	5.39
Na_2O	3.32	2.10	3.90	4.17	3.37
K_2O	2.00	1.66	3.29	3.21	2.54
H_2O+	1.49	1.40	.52	1.53	1.23
H_2O-	.33	.44	.20	1.55	.63
CO_2	.45	.13	.06	.83	.37
TiO_2	.74	.48	.45	.46	.53
P_2O_5	.40	.24	.23	.24	.28
S	.33	.....	.....	.....	.....
MnO	.12	.20	.04	.08	.11
BaO	.15	.25	.....	.....	.10
Total	100.20	99.89	99.97	99.97	99.92

* Pre-main intrusive sill.
** Pre-main intrusive dike.
† Main intrusive (pre-ore).
†† Post-main intrusive dike (post-ore).

A thick series of volcanic rocks overlies these beds. The late vitrophyric dikes cut the gravel and sand beds and were possibly the feeders

[2] General term used in the district for coarse granitoid to porphyritic rocks of pre-ore age which vary in composition from quartz diorite to quartz monzonite.

TABLE 2.—*Microscopic analyses of igneous rocks of the Pewabic mine area*

No.	P–20 (Pre-ore sill)		P–21 (Pre-ore dike)		P–19 (Main intrusive)		P–22 (Post-ore dike)	
Fabric	Porphyritic		Porphyritic		Holocrystalline		Porphyritic	
	No. Mineral	±%**	No. Mineral	±%**	No. Mineral	±%**	No. Mineral	±%**
Phenocrysts	1 hornblende (WD) 2 labradorite (An_{53})	15 25	1 hornblende 2 labradorite (An_{50}) 3 quartz	10 40	1 hornblende 2 andesine (An_{40}) 3 orthoclase 4 perthite 5 quartz 6 biotite (1)	5 40 27 10 15 3	1 hornblende (WD) 2 andesine (An_{30}) 3 orthoclase 4 quartz 5 biotite	3 25 5 −1 2
Groundmass minerals	3 quartz 4 hornblende (WD) 5 plagioclase 6 orthoclase	1 5 42 10?	4 quartz 5 hornblende 6 plagioclase 7 orthoclase	10? 5 5 30?			6 quartz 7 orth 8 plagioclase 9 glass	65
Accessory minerals	7 apatite 8 ilmenite (?) 9 magnetite (?)	−1 1 1	8 apatite 9 sphene	−1 −1	7 apatite 8 sphene 9 magnetite	−1 −1 −1	10 apatite 11 sphene 12 magnetite	−1 −1 1
Alteration minerals (young ← old)	10 ferro-tremolite (1, 2, 4)* biotite (2) leucoxene (8) pyrite (9, 10) siderite (?) (10) quartz (2)	25 −1 2 1 −1 −1	epidote (2) chlorite (2) quartz	5 −1 −1	sericite (3)	−1	biotite (1) magnetite (1) carbonate (1) chlorite (2) allophane (1, 2)	3
Total alteration		28		5		−1		3
Minerals in cracks	pyrite siderite (?) ferro-tremolite							
Ridgway color	hathi gray 35″″″, b		court gray 29″″″, GGY, f		mostly pinkish white		hathi gray 35″″″, b	
Specific gravity	2.732		2.735		2.701		2.645	
Per cent porosity	1.062		1.427		1.922		4.100	
Paragenesis (old → young)	hornblende labradorite hornblende plagioclase quartz orthoclase biotite ferro-tremolite pyrite quartz siderite (?)		hornblende labradorite quartz orthoclase epidote chlorite quartz		apatite andesine quartz orthoclase hornblende biotite magnetite sphene sericite			

WD Wholly Destroyed.

* Index numbers of minerals replaced.

** Estimated per cent.

of the volcanic rocks. The dikes and bedded volcanic rocks, closely related in composition, are generally more acidic than the pre-Eocene (?) rock. Regional correlations suggest that they are post-Eocene, possibly Miocene-Pliocene, but no accurate dating has been attempted. The age of the post-Cambrian pre-Tertiary igneous rocks is presumably Laramide.

Four specimens of igneous rocks from the Pewabic mine area were taken for chemical analysis. One is from the early Hanover or "marker" sill, another is from a pre-ore dike which cuts the sill but in turn is cut by the Hanover intrusive, another is from the Hanover intrusive, and the fourth from a post-ore vitrophyric dike. The chemical analyses of these specimens (Table 1), when compared with Daly's average analyses (Daly, 1933, p. 9-28), show a range in composition from near the average diorite through basic quartz diorite, quartz monzonite, to dacite. The microscope indicates approximately the same composition but shows free quartz in all slides including those of the early sills (Table 2, P-21 and P-22). In the early sill, however, the quartz has been partially resorbed. The classification is summarized as follows:

Group	Specimen No.	Near Daly's Average	Microscopic Classification
Early sill	P–20	diorite	hornblende quartz diorite
" dike	P–21	(basic) quartz diorite	hornblende quartz diorite porphyry
Hanover intrusive	P–19	" " monzonite	hornblende quartz monzonite
Late dike	P–22	dacite to quartz latite	hornblende dacite porphyry

The average of the four analyses falls between the average quartz diorite and granodiorite. The chemical and microscopical analyses show the close relationship of the four rocks. When the oxides and other data are plotted against relative age (Fig. 2), it is apparent that several changes in the probably common source magma occurred with time progression. The amount of silica, alkalies, and the total alkali : lime ratio increased with time while the amount of total iron, alumina, magnesia, lime, and the soda : potash ratio decreased. The specific gravity decreased slightly, the porosity increased, and the color bleached, due, of course, to the decrease in ferromagnesian minerals. The progressive increase in porosity may be due to capillary cracks developed more readily in the rocks, such as P-22, which cooled at the shallower depths. The other rocks are coarser grained and presumably were intruded at greater depths. P-22 looks more porous than the others and adheres to the tongue slightly. In the field the sill and dike rocks (P-20, P-21, and P-22) are at first difficult to distinguish. However, the sill rock has a fine-grained matted appearance owing to the fine needles of the labra-

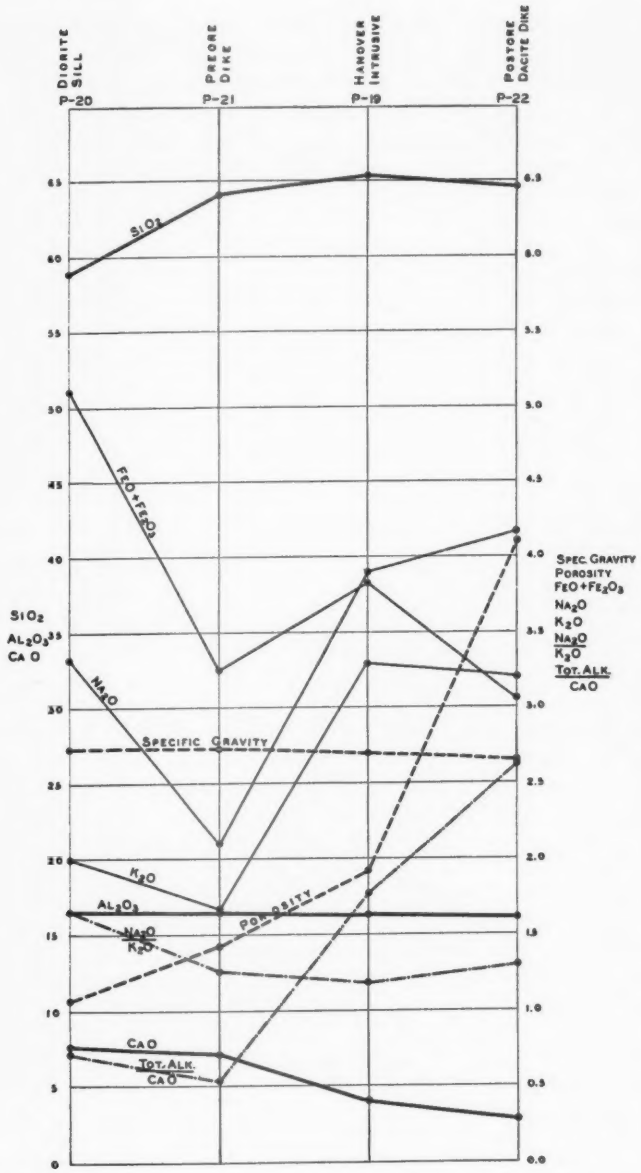


FIGURE 2.—*Relative age-composition curves for the igneous rocks*

dorite and hornblende, whereas the pre-ore dike rocks (P-21) are coarse textured and have large needles of black hornblende and phenocrysts of white feldspar. In certain parts of related dikes the hornblende is replaced wholly or partly by macroscopic crystals of biotite elongated with the C axis, and the feldspar may appear as large rhombs. The post-ore dikes are light gray, contain typical small shiny black crystals of biotite, and usually show a strong, fine platy flow structure parallel to their contacts.

The relative age of the sills, dikes, and the pyrometasomatism is revealed by relationships which can be traced on the surface in the Pewabic mine area (Pl. 2). The Hanover sill is cut by the pre-ore dike and in turn both sill and dike are cut and folded by the Hanover intrusive. The three rocks were later pyrometasomatized. The post-ore dikes which cut all the rocks described above are not pyrometasomatized; on the contrary, they contain breccia fragments of metamorphic silicates and ore which had been plucked from the walls by the magma.

STRUCTURE

FOLDS AND FAULTS

The structure of the Pewabic mine area is perhaps the most fascinating feature of the local geology. As the south lobe of the Hanover intrusive (Fig. 1) intruded as an ethmolithoid sheet, it thrust its walls aside to the east, south, and west, and this thrust was compensated by folding in an overthrust plate. The intensity of folding caused overturn (Pl. 2, section AA′). There seems to be no alternative interpretation for the thrust structure. The folding has been proved by detailed mapping of the extensive workings of the Pewabic mine and by about 25,000 feet of diamond drilling. A basal thrust fault is revealed at several places in the mine as a low-angle fault plane filled with breccia, gouge, or jasperoidized and/or silicified gouge. At a predictable line on the surface the fault plane outcrops as a 1- to 3-foot zone of jasperoidized gouge [3] and breccia. The fault is largely confined to the easily deformed Hanover limestone, and the attitude of faulted sections of this and adjacent beds indicates large offsets of the recognizable formations (Pl. 2, section DD′). It does not seem possible to interpret the fault as a regional structure because the extensive surface mapping and diamond drilling throughout the district (about 75,000 feet of diamond drill holes) do not indicate the structural form and stratigraphic discrepancies which should then exist. Diamond drilling and development in the neighboring Empire Zinc Company property, which occupies the south and west side of the Hanover

[3] Gouge, resembling brown hornstone, baked and partially jasperoidized and silicified by the metamorphism.

lobe (Fig. 1) of the intrusive, suggest a corresponding overthrust upper plate, the flat thrust fault occurring chiefly in the Hanover limestone. On the south side of the Hanover lobe, deep-level development in the Republic mine shows that the beds in the lower plate were nearly flat and unfolded and that the pronounced peripheral anticlinal form so prominent on the surface disappears with depth.

In the Pewabic area, section AA′ (Pl. 2) suggests that the thrust fault disappears in the peripheral anticline, the necessary compensation having occurred by rock flow in the Hanover and Middle Blue limestones which are thickened and thinned. That the limestones actually flowed is proved by the marked flow structure outlined by the white chert beds which, though normally parallel to the bedding, are folded, overturned, and pulled apart in many places. The flowage usually resulted in thickening in the anticline and thinning in the syncline (Schmitt, 1935, p. 38, fig. 3). In 1937 two diamond drill holes were put down in the Hanover intrusive. They encountered metamorphosed sedimentary rock at elevations predictable by the extrapolation of the basal thrust fault (Pl. 2, section AA′).

Subsequent to the thrusting, northeast-striking high-angle normal faulting occurred with small slips, but with important ore-localizing effects. This faulting was probably regional in origin because it is regional in distribution and attitude as contrasted with the thrusting which, so far as is known, is local and related to the Hanover intrusive.

The normal faults, whose dip is generally more than 70 degrees, are shown on Figure 3, Plate 2, a and Plate 2, section AA′. They vary in age from pre-silicate metamorphism to post-silicate metamorphism and to post-ore and even post-middle (?) Tertiary dikes. The northeast faults fall into five zones designated Alpha, Beta, Gamma, Delta, and Epsilon (Pl. 2, a). Alpha and Epsilon contain post-ore dikes but do not localize ore. The others are ore-bearing and give their names to the three principal ore zones in the mine—Beta, Gamma, and Delta. They localize silicates (Fig. 3) and contain jasperoidized gouge, and individual sympathetic fractures cut garnet and localize veinlets of sphalerite in the garnet. Apparently some of the fractures of the zone were reopened in post-ore time for they are loci for deep though quantitatively unimportant oxidation. One of the fractures appears to offset a post-ore dike (Fig. 3) and so may represent movement as late as post-middle (?) Tertiary. The maximum slip on the ore-bearing fault zones is about 15 feet.

Several post-ore northwest-striking faults, with slips up to 20 feet, are known in the northeastern part of the mine. They generally localize post-ore dikes (Fig. 3) and offset the ore in many places but have not greatly interfered with development and mining.

DIKES

All dikes in the Pewabic area stand at high angles. The pre-ore dike antedates the Hanover intrusive because it is cut by an apophysis of the intrusive (Pl. 2). The eastward thrust of the Hanover intrusive

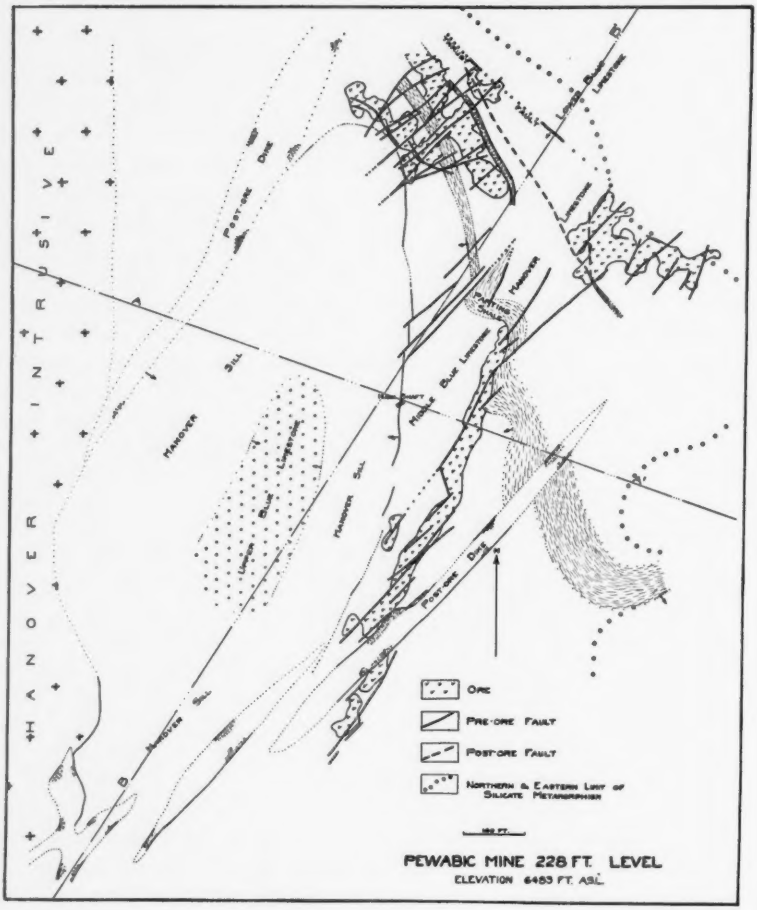


FIGURE 3.—*Geology of the 228-foot level*
Location of sections AA′ and BB′ shown.

in this area has bent the dike convexly eastward (Pl. 2, and Figure 1). The westward dip of the dike decreases with depth because of the drag near the thrust fault.

Most post-ore dikes cut ore bodies, always on planes of faulting, which commonly have reopened and brecciated the dike. They range in strike, not having the nearly constant northeast strike of the pre-ore dikes, and range greatly in thickness along the strike. This tendency is a characteristic not shared by the pre-ore dikes in the Hanover district (Fig. 3).

MINERALS

The hypogene and some of the supergene minerals of the Pewabic area are described as follows:

ALLOPHANE (Table 3) occurs as a minor alteration product of hornblende and andesine in the post-ore dacite porphyry but is not known in the earlier dikes.

ANDESINE (Table 3) is the plagioclase of the later dikes. It is commonly euhedral and occasionally broken as though by movement subsequent to crystallization. The anorthite molecule ranges from 40-30 per cent.

ANDRADITE is the common garnet of the district and constitutes about 58 per cent by volume and 60 per cent by weight of the metamorphic zone at the Pewabic mine. The chemical and spectrographic analyses of a garnet specimen from the Hanover limestone (Tables 4 and 5) show that it is nearly pure andradite. There is, however, 0.98 per cent MnO. Optical data from the same specimen are given in Table 6. The shaly blue limestone above and below the Hanover limestone generally contains a darker, finer-grained garnet which possibly is aluminous. Both kinds of garnet commonly show birefringence.

ANTIGORITE (?)—Many slides cut from the ore show antigorite (?) (Table 6) as veinlets cutting the salite. These veinlets also contain a carbonate which is probably calcite. The Central Mining District is a "serpentine-talc-crysotile-magnetite" province. Much "serpentine" and magnetite occur at Fierro. Small veinlets of crysotile cut the "serpentine". The "serpentine" at Fierro may be a direct alteration product of dolomite, for no suggestion of earlier pyroxene has been noted in or near the massive "serpentine". In the Pewabic mine, however, it is normally an alteration product of salite.

APATITE occurs in relatively large euhedral crystals in all the intrusive igneous rocks studied. The writer has seen none associated with the pyrometasomatic minerals in the Pewabic area, but a small amount occurs in the metamorphic zone west of the Hanover lobe of the intrusive, and well-formed crystals are found in the north pit at the Santa Rita copper mine associated with specularite in a vertical zone of brecciated granodiorite.

BIOTITE high in iron (Table 3) is characteristic of most of the "cross-cutting" igneous rocks. It shows the typical pseudo-hexagonal form in the main Hanover intrusion with the books elongated as much as three-eighths of an inch parallel to the C axis. The microscope suggests that it progressively approaches a pyrogenic character with the decrease in age of the intrusives. Thus, in the early sill rock (Table 2, P-20) jointly with ferro-tremolite it is apparently a deuteric alteration product of labradorite. Hornblende is the sole primary ferro-magnesian mineral. In the later Hanover granodiorite intrusive (Table 2, P-19), it appears as an apparently deuteric alteration product of hornblende. In the post-ore volcanic dikes (Table 2, P-22), however, it is both pyrogenic and deuteric, for although some is replacing hornblende some is euhedral, broken and bent, as though it had crystallized in the magma before movement had stopped.

CARBONATE—The chief carbonate is, of course, the calcite in the limestone and marble. Analyses of coarsely crystalline calcite associated with coarse garnet and ore minerals are given in Table 4. Minor carbonate has the optical characteristics of siderite. Where massive silicates are in contact with white marble a zone in the marble at the contact, averaging about 2 inches in width, has been altered to a manganiferous, possibly ferriferous, carbonate which, when weathered, develops a brownish coating. No observable textural change accompanies this compositional change.

CHALCOPYRITE occurs as the usual blebs in the sphalerite throughout the mine. In the lowest part of the mine, on the 370 level, in the area nearest the Hanover intru-

TABLE 3.—*Optical characteristics of pyrogenic minerals of the Pewabic mine area*

No.	Mineral	Pleochroism			Index of refraction			Birefringence	Optic character	2V	Extinction	Dispersion	Remarks	Twinning
		X	Y	Z	X	Y	Z							
P–19	Biotite	Light brown		Seal brown					−	0 or small	0			
P–22	Andesine							.01 ?	−	±80			An 40	albite carlsbad pericline
P–20	Labradorite					1.54			+	±80	with 010 30°			albite pericline
P–21	Hornblende	yellow green	grass green	bluish green				.02	−	±60	with elongation 27°			
P–26	Sericite	colorless	colorless	colorless		1.55		.026 ?	− ?				Elongation (+)	
P–36	Augite			light gray				.022 ?	+	±60	$Z \wedge C = 40°$	$P \wedge Z > V \wedge Z$		
P–37A	Apatite					1.62		.001	−	0			Elongation (−)	
P–22	Allophane	pale brown				1.54		0					Amorphous	

sive, several tons of massive chalcopyrite has been found in the zinc ore bodies. Because pyrrhotite and magnetite are also more abundant here, it is believed that the chalcopyrite favors the areas of higher temperature of deposition. Masses of more or less pure chalcopyrite associated with abundant magnetite were found on

TABLE 4.—*Chemical analyses of hypogene minerals of the Pewabic mine area*

(R. B. Ellestad and A. Willman, analysts; direction, F. F. Grout)

	Andradite P–31	Salite P–33	Salite* P–5B	Ilvaite P–32	Calcite**	Altered shale (98% epidote) P–29	Thuringite†
SiO_2	36.27	52.20	48.52	29.16	.....	36.48	31.80
Al_2O_3	.51	.26	1.67	.23	.....	21.37	19.99
Fe_2O_3	29.97	.74	1.86	20.28	.....	12.46	5.21
FeO	.57	14.84	18.56	31.06	1.39	1.32	25.20
MgO	.14	7.50	4.11	.67	.17	.34	4.75
CaO	31.61	22.19	21.71	13.29	53.50	21.77	1.17
Na_2O	.....	.....	.....	.....	.....	.06	.06
K_2O	.....	.....	.....	.....	.....	.35	.62
H_2O	0.9	.31	.75	2.19	.....	2.07	9.09
H_2O		.07	.24	.07	.....	.33	.....
CO_2	.....	.....	.....	.....	41.80	1.21	.84
TiO_2	.01	.01	.08	.01	.....	1.66	.84
P_2O_5	.....	.....	.....	.....	.....	.05	.40
MnO	.98	2.17	2.66	3.13	1.87	.27	.56
Total	100.15	100.29	100.16	100.09	98.73	99.74	99.98

* Darker salite.

** W. B. Fisher, Deming, New Mexico, Analyst.

† Recalculated from analysis P–26.

the 300 level of the Republic mine (Fig. 1) near the main Hanover intrusive. Magnetite with as much as 5 per cent copper as copper carbonate was mined in the Hanover zinc mine in a zone about 100 feet wide bordering the main intrusive contact. Presumably the copper was derived from chalcopyrite. At the Philadelphia mine, within the south lobe of the Hanover intrusive (Fig. 1), coarse chalcopyrite occurs with massive magnetite and coarse ferro-tremolite, epidote, quartz, orthoclase, and other minerals. These minerals are localized by faults and xenoliths of sedimentary rocks in the main intrusive. The Pewabic mine zinc concentrates averaged .01 per cent Cu during 1933 and 1934, and the partial analysis of a sphalerite specimen, which under the microscope appears uncontaminated, gave by chemical analysis .04 per cent Cu.

CHLORITE—Two important occurrences of thuringite (Tables 4 and 6) are known to the writer in the Hanover intrusive metamorphic area. This chlorite is an alteration product of the same pre-Hanover intrusive dike and in two different areas replaces the entire dike as a mass of nearly pure thuringite. The large percentage of ferrous iron and combined water in this mineral suggests a late stage of alteration connected with the ore mineralization, later than the main garnet-epidote stage.

EPIDOTE—An epidote varying slightly from a 12.00 per cent content of ferric oxide, as indicated by chemical analyses and microscopical identifications (Tables 4 and 6), is common. Generally it is an alteration product of the aluminous rocks only, and consequently the altered dikes and shales within the metamorphic zone may be readily mapped by their epidote content distinct from the altered limestone which alters to andradite. At the Philadelphia shaft, southwest of the Pewabic shaft (Fig. 1), epidote in fine tufted crystals up to an inch long is associated with large actinolite crystals, as much as 6 inches long, and large orthoclase, quartz, magnetite, and

TABLE 5.—*Roughly quantitative quartz spectrograph analyses of certain hypogene minerals of the Pewabic Mine*

(By John Herman Laboratory)

	Andradite P-31	Salite P-33	Salite* P-5B	Ilvaite P-32	Sphalerite P-6A	January 1935 Zn concentrates
Si	10.0+	10.0+	10.0+	10.0	1.0	1.0 0.1
Fe	10.0+	10.0+	10.0+	10.0	10.0+	10.0
Mn	1.0	0.1	1.0 0.1	10.0	0.01	0.1 .01
Ca	1.0 .01	1.0 .01	0.1	1.0 .01	0.1	0.1 0.01
Mg	0.1	1.0 .01	1.0 .01	1.0 0.1	0.001	.01 .001
Zn	0.1 .01	.01	.01	.....	10.0	10.0
Cu	0.01 .001	.001 .0001	.01	.....	.001	0.1
Ba	0.001	.....	0.001	0.01 0.001	.....	.....
Ni	.....	.01	.01 .001	.....	.....	.....
Co	.....	0.001	.....	.....	.....	.....
B	.....	0.001	0.001	.....	0.001	.....
Al	.....	.....	.01 .001	.....	.....	.01
Pb	.....	.....	.001	.....	.001	.01
Sn	.....	.....	.001	.....	.....	.....
Cr	.....	.....	0.001	.....	.....	.....
Ag	.....	.....	0.0001	.....	.....	.001 .000
Cd	.....	.....	.....	.....	0.1 0.01	0.1 .01
V	.....	.....	.....	.....	.01 .001	.....
Ti	.....	.....	.....	.....	.001	.....
Bi	.....	.....	.....	.....	.....	.01

* Darker salite.

chalcopyrite crystals. These minerals mark a focus of alteration well within the Hanover intrusive. Both intrusive rock and xenoliths (?) of sedimentary rocks are replaced. It is probably a late magmatic alteration related to the deuteric and/or pegmatitic stage but may be a link with the pyrometasomatic stage because many of the same minerals though finer grained occur in the main pyrometasomatic zone.

GALENA is even rarer than chalcopyrite in the Pewabic mine. The temperature was probably generally too high for its deposition. Some coarse galena occurs in the Beta zone which is the zone farthest away from the hot center previously mentioned. Galena is rare also in the part of the contact zone owned by the Empire Zinc Company where it occurs, for the most part, at the outer edge of the metamorphic zone; some occurs in 1- to 3-foot jackets, which envelop many of the sphalerite ore bodies.

The occurrence of galena must be watched because the premium commanded by the Hanover zinc concentrate is due to low lead content. In the Empire Zinc Company property galena when found in the stopes as coarse masses was frequently removed in powder boxes and thrown away. The Peru mine concentrates have shown only a trace of lead for a 2-year period. The spectrograph showed .01 per cent in the January 1936 concentrates. For several years the Empire Zinc Company had a ratio of about 300 to 1 of zinc to lead.

GARNET—Andradite is the common garnet and has already been described. As was discussed, aluminous garnet may occur in the impure limestones.

HEMATITE as specularite occurs in small quantities in the matrix of quartz which is commonly intersertal to massive coarsely crystallized andradite. Possibly after the lime had been exhausted in the formation of andradite from the limestone the same mineralizing fluids could only deposit quartz and ferric oxide in the remaining spaces.

HORNBLENDE in the igneous rocks of the area tested is always the same species (Table 3). The angle made by the slow ray with the axis C is 27 degrees, which seems abnormally high.

Much of the hornblende in the Hanover sill rock was altered to ferro-tremolite. (*See* Table 2.) The hornblende in the mainly cross-cutting rock, however, was partly altered to biotite. As a result, the Hanover main intrusive in places is a biotite monzonite, and all ratios of hornblende to biotite are present. Likewise, the composition of a single pre-ore dike may vary from hornblende porphyry to biotite porphyry. It is not clear whether this transformation of the ferromagnesian mineral occurred in the magma chamber or by deuteric alteration after emplacement. The principal focus of the biotite is the center of the intrusive with gradations to a border hornblende phase particularly noticeable in the Hanover lobe of the main intrusive.

Some dikes are wholly biotite porphyry and may have been intruded as such. For example, the biotite crystals in the post-ore dikes are bent and so were formed in the magma.

ILVAITE is one of the chief gangue minerals of the Pewabic mine. The only other known important occurrence in the district is on the west side of the Hanover intrusive at N5000, W3600, and El. 6440 (Fig. 1). The chemical, spectrographic, and optical analyses are given in Tables 4. 5 and 6. Many well-formed crystals occur at the two places mentioned. Much of the Pewabic mine ore, particularly near the intrusive contact, contains abundant ilvaite. Ilvaite is rarely found at a distance from sphalerite.

LABRADORITE (An_{50}) is the chief plagioclase in the pre-ore Hanover sill and pre-ore dike (Table 2, P-20 and P-21). The plagioclase of the younger rocks is andesine.

LEUCOXENE occurs in the main Hanover intrusive rock and was probably derived from ilmenite.

"LIMONITE" occurs mainly in the gossan outcrop of the Pewabic ore lens (Pl. 1c) and was derived chiefly from the oxidation of pyrrhotite and ferriferous sphalerite.

MAGNETITE is common in the ore but, unlike ilvaite, may occur a distance from sphalerite. It contaminates much of the sphalerite as vermicular inclusions too small to be detached by commercial grinding. The mill men call this magnetic sphalerite. Much of the magnetite is pure enough to be iron ore, and some was shipped in 1936 by the Empire Zinc Company from ore bodies in the andradite zone. At the surface much of this magnetite is martite, but no martite is known at depth. The writer believes the oxidation has occurred through bathing by ferric sulphate solutions derived from the oxidation of the associated pyrite, pyrrhotite,

TABLE 6.—*Optical characteristics of certain hypogene minerals of the Pewabic mine*

No.	Ridgway Color	S. G.	Pleochroism			Index of Refraction*			Birefringence	Optical Character	2V	Extinction	Dispersion	Remarks
			Y	Y	Z	X	Y	Z						
P–31 Andradite	Dresden brown	3.838					±1.9							Coarse crystals
P–33 Salite (light)	Lincoln green	3.447	Yellowish green	Almost colorless	Bluish green	1.687	1.692	1.721	.034	+	±60	Z∧C=54°	Z∧C P>V	Coarse crystals
P–5B Salite (dark)	Deep slate-olive	3.502	Very light green			1.693	1.690	1.715	.025	+	±60	Z∧C=38°	Z∧C P>V	Fine crystals
P–32 Ilvaite	Black	4.009	Greenish brown		Dark green to opaque	>1.78								Coarse crystals
P–17 Epidote	Serpentine green	3.355	Colorless	Light lemon yellow	lemon yellow	1.715 (±.003)	1.742 (±.003)	1.760 (±.003)	.045	−	±85	Z∧001=27°		Coarse crystals
P–7A Ferrotremolite	Dark ivy green		Colorless	Colorless	Bluish green	1.636		1.665	.029	−	±80	Z∧C=18°		Radiating
P–15 Thuringite	Dark grayish olive		Colorless	Weak bluish green	Bluish green	1.642 (±.003)	1.655 (±.003)	1.656 (±.003)	.014	−	very small	Z∧base=±4°		Radiating
P–9 Antigorite (?)			Pale-yellowish-green ±½, 21, i						.025?			Parallel		Alteration of salite
P–12 Talc	Hathi gray		Light bluish green	Colorless	Light yellowish green	1.557 (±.002)	1.600 (±.002)	1.605 (±.002)	.048	−	±15	±2½°		Micaceous
P–36 Marialite			Colorless	Colorless	Colorless		±1.55		.006	−	0	Parallel		Prismatic
P–27 Unknown in shale			½, 21, i yellow		Dark yellow ½, 21, k	1.63		1.65	.025?	−	±85	Z ‖ with elongation		Foliated, radiating
Wollastonite	White		Colorless	Colorless	Colorless	1.620 (±.001)	1.625 (±.001)	1.629 (±.001)	.009	−	±40	‖		Radiating

* Indexes determined by use of heatable stage.

and ferriferous sphalerite. In the Pewabic mine, talc (Table 6) is a nearly universal contaminant of the magnetite.

MARIALITE has been identified in one slide of the altered main intrusive associated with deuteric (?) augite (Table 6). Insofar as known, this augitization and scapolitization have affected only one small area of intrusive in the Pewabic area immediately adjacent to the sedimentary contact (N4725, E1080, El. 6747) and are associated with aplite dikes and garnet and epidote veins which cut the aplite. Other halogen-bearing minerals, namely fluorite and vesuvianite, occur on the west side of the Hanover intrusive about 2000 feet west of the main garnet zone at the outer limit of the hornstonization of the shales.

ORTHOCLASE occurs in the igneous rocks of the area, mainly in the groundmass. No microcline has been seen, but there is perthite in the main intrusive monzonite. Coarse orthoclase, associated with large crystals of quartz, magnetite, ferro-tremolite, epidote, and chalcopyrite, is found at the Philadelphia shaft.

PYRITE, though common, is not abundant. As usual it is ubiquitous but in the Pewabic mine the total amount is less than that of pyrrhotite. In places the microscope shows it to be an alteration product of pyrrhotite, magnetite, and actinolite.

PYRRHOTITE is one of the chief gangue minerals of the "hot center" part of the ore-producing area where it is about as abundant as ilvaite. It is associated with a minor amount of chalcopyrite, but neither of these minerals is found in the Pewabic mine in notable quantity outside of the "hot center" area. Pyrrhotite is abundant in the "Pewabic ore lens" (Pl. 2) which lies within the "hot center" area.

QUARTZ, though widespread, is quantitatively unimportant. Its association with specularite and andradite has been described under hematite. It also occurs in veinlets associated with sphalerite, carbonates, and specularite, filling pre-sulphide post-garnet faults and fractures.

SALITE OR SAHLITE (Bayley, 1917, p. 373; Winchell, 1933, p. 225), a mineral intermediate between diopside and hedenbergite, is the common pyroxene of the district. Two closely related varieties are recognized in the Pewabic area. (*See* Tables 4, 5, and 6, specimens P-33 and P-5b for chemical, spectrographic, and optical characteristics.) One is the coarsely crystalline variety with light-green bladed crystals up to 6 inches long; the other is a fine crystalline, usually matted, dark-green variety which is slightly higher in ferric and ferrous iron and has a greater specific gravity. Under the microscope the second has less pleochroism and a lower birefringence than the first, but Nm is about the same for both. In the second, the slow ray makes a smaller angle with the C axis. The second, microscopically, shows numerous small inclusions, and the spectrograph shows more impurities (Table 5) than the first. A considerable variation in the pleochroism in both varieties implies a variation in composition common to pyroxenes. Manganese occurs in all the pyroxene of the district. A specimen analyzed by Charles Park in 1928 yielded 20 per cent Mn.[4] After a few months of weathering, all the salite develops a brown to jet black coating of manganese oxide, the depth of color, other things being constant, depending on the time of exposure. Nearly all the minerals of the area, including sphalerite, contain manganese and develop brown to black coatings on exposure. (*See* Tables 4 and 5.)

In the metamorphic aureole nearest the intrusive on both the Peru Mining Company and Empire Zinc Company properties, the coarse salite (locally called "hornblende" or hedenbergite) almost never occurs without associated sphalerite. In only two areas does it occur without sphalerite—on the west side of the main intrusive at about N7100, W4400, El. 6600, and on the east side at about N6200, E2600, El. 6800. It was early recognized as a guide to ore. In the gossans the characteristic bladed, radiating crystal texture is preserved as a dark limonite. The sphalerite weathers to a limonite of a lighter shade. This makes possible a close estimate of the former zinc content of a given gossan.

The coarse crystalline salite is rarer and the fine crystalline variety more common at the Pewabic mine than at the Empire Zinc Company mines. The dark variety is confined chiefly to the areas nearest the intrusive and is characteristically associated with pyrrhotite, ilvaite, and chalcopyrite.

[4] This is apparently the Johannsenite described by Schaller (1938).

SERICITE—Some sericitic alteration accompanied by quartz, calcite, and pyrite has affected the post-ore dikes and, on the property of the Empire Zinc Company, a few of the pre-ore dikes. In the Chino, Ground Hog, and other mines where lower temperature conditions may have prevailed, sericitization and chloritization of the older igneous rocks are common. Epidote is, however, associated with the chlorite.

SIDERITE—A small amount is found in the Pewabic area, usually as veinlets cutting other minerals. It is the only mineral apparently related to the ore mineralization which occurs in the marble at the surface above the easternmost blanket ore body in the Pewabic mine. It occurs in veinlets in marble at the surface over a blanket ore body west of the Hanover intrusive lobe where no other surface indications were noted. Siderite is more common in the Ground Hog mine where ore deposition may have taken place under cooler conditions.

SPHALERITE is the chief ore mineral of the pyrometasomatic zone, and its freedom from lead makes it especially desirable for smelting. A specimen of macroscopically clean sphalerite was examined in polished section. The only impurity noted was bleboid chalcopyrite visible at 450 diameters. The chemical analysis of this specimen by W. B. Fisher of Deming, New Mexico, was as follows:

Zn	59.4
Pb	Trace
Cu	.04
Fe++	6.0
Fe+++	.1
S	32.0
Insol	1.0
	98.54

The spectrographic analysis of the same material is given in Table 5. Some large bodies of massive sphalerite assayed 50 per cent Zn, and with lead never more than .1 per cent. The cadmium content is usually reported as a trace. Vanadium (Table 5) is of course a common impurity of sphalerite and probably accounts in part for some of the vanadium minerals common to the oxidized zone of zinc-lead ore bodies in the Central Mining District.

SPHENE occurs as characteristic euhedral form in all the intrusives younger than the epoch of sill intrusion.

TALC is found in the Pewabic mine, usually with magnetite. The optical characteristics are given in Table 6. The slow ray appears to make an angle of about 2½ degrees with the basal cleavage so, as suggested by Rogers and Kerr (1933, p. 281, 282), the mineral is probably monoclinic.

THURINGITE has been discussed under chlorite.

TREMOLITE—Ferro-tremolite is a common mineral usually occurring as an alteration product of the hornblende of the igneous rocks—an alteration which may be complete insofar as the hornblende is concerned. Some is found in the Pewabic mine as an alteration product of salite. Judging from the bent acicular forms which it takes where squeezed by movement, this is probably the mineral from which the "mountain leather" and actinolite asbestos of the Hanover area are derived.

WOLLASTONITE—One specimen of wollastonite (Table 6) was found at N5140, E2065, El. 6778, at the outer edge of the metamorphic zone. It formed at the boundary of white chert with marble. The metamorphosing agents would be expected to be deficient in iron, and reaction dependent on heat and the elements at hand. Abundant wollastonite is reported from Fierro and recently has been found in abundance on the south part of the Peru Mining Company's property near Santa Rita.

UNKNOWN MINERAL—An unidentified mineral whose optical characteristics are given in Table 6 (P-27) was noted in cracks in what macroscopically appears to be hornstone but is really a layer of fine-grained impure quartzite in the Parting shale. Most of the mineral cuts the quartzite as veinlets, but some is disseminated. When dissolved with acid it gives strong tests for Mg and Fe but none for Ca, SO_4, CO_3,

TABLE 7.—*Chemical analyses showing progressive metamorphism of certain rocks of the Pewabic mine area*

(R. B. Ellestad, A. Willman, analysts; Direction, F. F. Grout)

	Hornstone (shale) P-27	Epidote P-29	Dike P-21	Epidote P-23	Chlorite P-26	Limestone	Andradite P-31	Salite P-33
SiO_2	76.29	36.48	64.15	42.21	38.95	.....	36.27	52.20
Al_2O_3	10.54	21.37	16.45	16.06	18.13	.....	.51	.26
Fe_2O_3	1.67	12.46	1.13	10.55	5.07	.....	29.97	.74
FeO	4.71	1.32	2.12	3.37	21.50	.....	.57	14.84
MgO	1.19	.34	2.05	2.51	4.06	.....	.14	7.50
CaO	.10	21.77	7.09	19.33	2.09	.....	31.61	22.19
Na_2O	.17	.06	2.10	.03	.05	.....	.....	.....
K_2O	1.45	.35	1.66	.04	.53	.....	.....	.....
H_2O+	2.63	2.07	1.40	1.78	7.76	.....	.09	.31
H_2O-	.17	.33	.44	.19	.25	.....	.....	.07
CO_2	.....	1.21	.13	1.03	.....	.....	.....	.....
TiO_2	.93	1.66	.48	.59	.72	.....	.01	.01
P_2O_5	.04	.05	.24	.25	.34	.....	.....	.....
S	.....	.....	.....	.07	.....	.....	.....	.....
MnO	.11	.27	.20	.48	.48	.....	.98	2.17
BaO	.....	.....	.25	.....	.....	.....	.....	.....
ZnO	.....	.....	.....	1.4	.....	.....	.....	.....
Total	100.00	99.74	99.89	99.89	99.93	.....	100.15	100.29
S. G.*	2.790	3.355	2.735	3.275	3.018	2.715	3.838	3.447

* Specific gravity corrected for H_2O at 4° C.
Minerals cleaned by use of a magnet, acid and s-tetra-bromoethane.

or PO_4. The analysis (P-27), when recalculated after subtracting SiO_2 (quartzite) and the titanium (probably residual titanite), gives:

Al_2O_3	47.50
Fe_2O_3	7.52
FeO	21.22
MgO	5.37
K_2O	6.54
H_2O	11.85
	100.00

METAMORPHISM

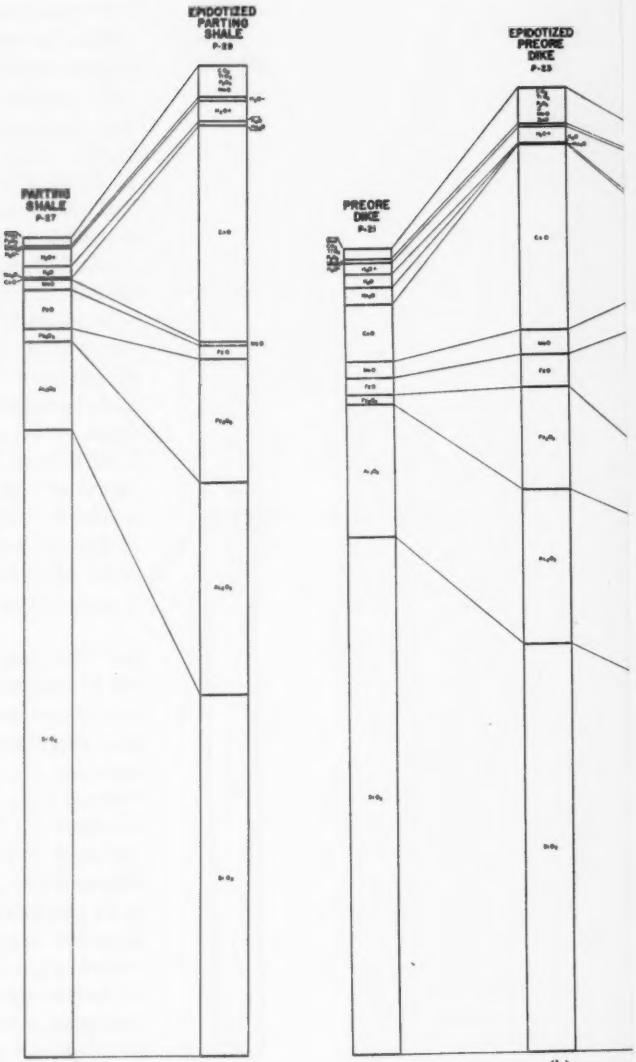
The Hanover intrusive is surrounded by an aureole of metamorphic rocks, chiefly replacements of limestones and shales but partly of dikes and sills. The original rocks at the Fierro end of the Hanover intrusive were predominantly the magnesian limestones and dolomites of the lower Paleozoic section below the Percha shale, but the rocks at the Hanover end of the intrusive were dominantly the nonmagnesian shales and limestones of Carboniferous age which lay above the Percha shale. The

difference in magnesia is shown by analyses (Table 9) of the dolomites and limestones from specimens and samples collected over a period of years.

The great differences in the original composition of the rocks, particularly the higher magnesia in the older Paleozoic rocks at Fierro and the higher alumina in the younger rock at Hanover, apparently have caused the striking differences in the occurrence and proportions of the metamorphic minerals in the two areas. Serpentine, wollastonite, and magnetite predominate at Fierro whereas garnet, epidote, and sphalerite are rarely found except in the few uneroded remnants of Carboniferous rocks. At Hanover, garnet, epidote, hornstone, sphalerite, ilvaite, and salite predominate, and magnetite, though common, clearly avoids the shale and shaly limestone. The lack of magnesia apparently did not inhibit its deposition. Serpentine and wollastonite are rare at the surface where only Carboniferous and Cretaceous rocks outcrop, but deep exploration in the Republic mine disclosed large quantities of serpentine and magnetite below the Percha shale with garnet, epidote, and sphalerite, so abundant above the shale, largely absent.

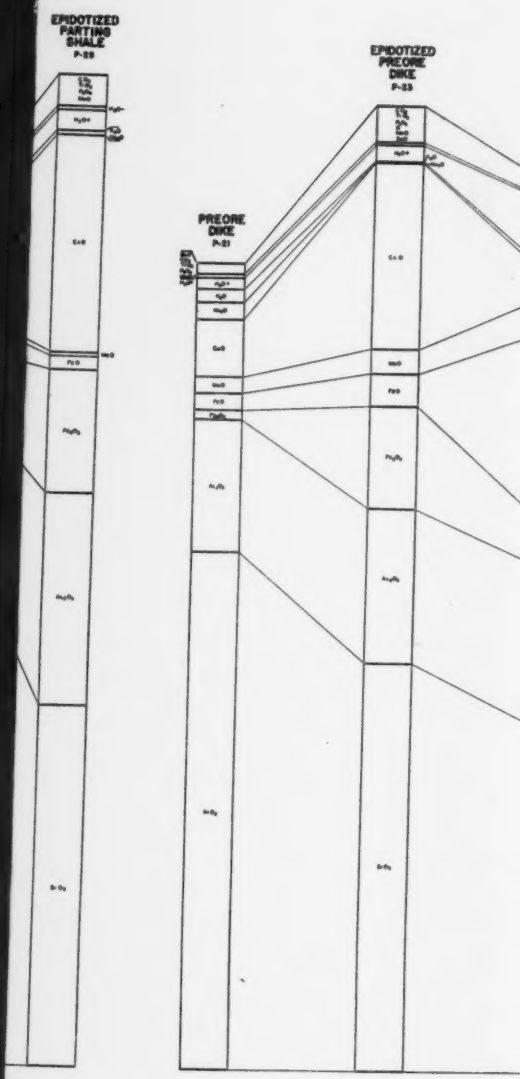
Some of the metamorphic minerals are strikingly confined to certain host rocks. Epidote invariably selects aluminous rocks, shale, and shaly limestone. Garnet, except in the most intensely metamorphosed areas, avoids the igneous rocks and shale and selects limestone or slightly shaly limestone. The coarse pure marble derived by metamorphism from the Hanover crinoidal limestone is replaced by coarse pure honey-yellow andradite (Tables 4, 5, and 6), but never by epidote. The less pure limestones are replaced by finer-grained darker, possibly aluminous, garnet and the resulting mass looks "dirty" compared with the massive pure garnet of the Hanover limestone. This selection of certain beds by epidote and garnet aids greatly in mapping the structure. The coarse salite is always found in the purer limestones, but ilvaite and fine dark salite are also found in the shaly limestones. The best ore favors the purer limestones, i.e., the Middle Blue and Hanover limestones. Chlorite and sericite are found only in the igneous rocks.

The composition-alteration diagrams (Pl. 1, a, b, and c) were compiled from the chemical analyses and specific gravities given in Table 7. Constant volume was assumed, and the diagrams were plotted in kilograms per cubic meter. The metamorphism of the Parting shale which is typical of the local shales is shown diagrammatically in Plate 1, a. Analysis P-27 was of a hornstonized phase of this shale which is probably of almost the same composition as the wholly unaltered shale except that about 2000 feet beyond the metamorphic zone the latter contains carbon and/or peat from numerous plant remains. The diagram suggests a



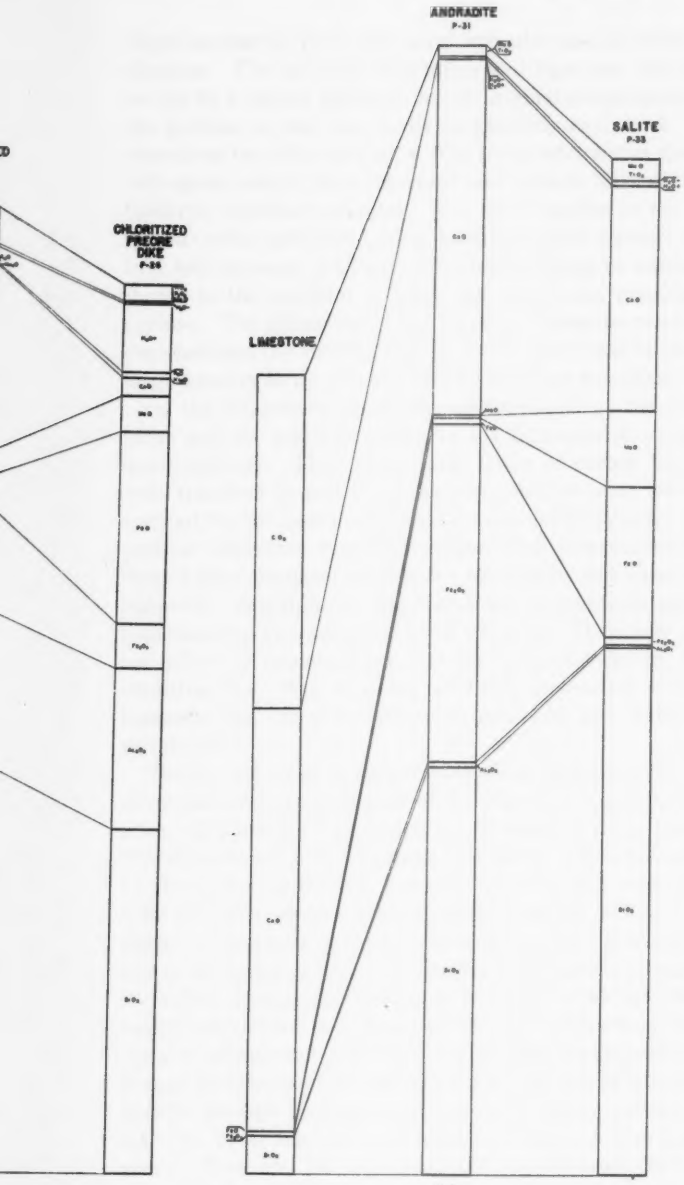
(a) (b)

METAMORPHISM OF T



(b)

METAMORPHISM OF THREE



(c)

OF THREE TYPICAL ROCKS

large increase in lime and ferric iron, decrease in silica, and increase in alumina. The increase in alumina and lime and decrease in silica may be due to a lateral variation in the original composition of the shale, but the increase in iron can hardly be similarly explained. Plate 1, b, which represents the alteration of a dike to epidote, shows that lime and ferric iron again seem to have increased and silica to have decreased. Alumina, however, remained constant. The epidotization of the aluminous rocks, igneous rocks, and shale, then, mainly implies increase in lime and ferric iron and decrease in silica. The chlorite stage of alteration of the dikes shown in the diagram is much less important quantitatively than the epidote. The alteration of the Hanover limestone to andradite (Pl. 1, c) demonstrates the striking though well-known type of alteration involving large gains in ferric iron and silica and losses in carbon dioxide.

Of the alterations given, the epidotization of the shales and igneous rocks and the andraditization of the limestone are the most important quantitatively. They imply large losses of carbon dioxide and gains in iron, transfers to and from the metamorphic zone which cannot be accounted for by local exchanges between unlike primary rocks. The composition-alteration diagrams suggest that importation and/or exportation of silica occurred, or that the local gains and losses of different rocks balanced. Individually, the limestones undoubtedly gained silica by replacement by andradite and other silicates. The shales and igneous rocks lost silica by epidotization, and the igneous rocks by epidotization and chloritization. But, in order to obtain a measure of the net gains and losses for the entire metamorphic zone (Pl. 2b) Table 8 was compiled as follows:

The Pewabic mine area, covering about 12 acres, is the most thoroughly developed and explored part of the Hanover section of the metamorphic zone. Exploration was largely by diamond drilling (about 25,000 feet). The detailed mine development was partly accomplished by the stoping, *i.e.*, by following the ore, but some drifting and cross-cutting have been done as deadwork to reach ore blocked out by drilling. The structure of the area, the limits of the metamorphism, and the distribution and quantity of the minerals are so thoroughly determined by extensive exploration and detailed mapping (scale usually 1 inch = 40 feet) that a good opportunity was offered to compile an accurate estimate of the composition of the metamorphosed zone before and after metamorphism, assuming no change in volume which the writer believes can be proved. Four vertical parallel geologic sections normal to the intrusive contact, of which section AA′ (Pl. 2) is one, were compiled and checked over a period of several years. They are 300 feet apart and cut the most thoroughly developed part of the mine. The average per cent of area of the original rocks cut

by the sections and of the metamorphic minerals replacing the original rocks should give a close approximation of the relative volume per cent of the original rocks and metamorphic minerals on the Pewabic mine side of the Hanover metamorphic zone. The areas of the original bedded rocks and cross-cutting igneous rocks were measured with a planimeter. The areas of the minerals were obtained partly by direct measurement, for there are blocks of ground which are nearly pure garnet, epidote, and magnetite, and partly by averages based on roughly quantitative underground estimates of the minerals.

Chief proof of constant volume during metamorphism is given by structural conditions. First, the form of the broad folded structures encircling the Hanover intrusive is logically explained as the result of peripheral thrust by the intruding magma. The force and confinement factors were adequate to induce flowage which thickened the anticlinal axes and thinned the limbs and the synclinal axes (Pl. 2). The metamorphism, however, engulfs the resulting structure without changing its form to any discernible degree. Second, when the bedded limestones and shales abut dikes along which metamorphism is localized there is no apparent thickening or thinning of the beds at the metamorphic contact. As the structure in the Hanover area is intimately known through extensive diamond drilling and other development work any change greater than about 5 per cent in the thickness of beds would be revealed where metamorphosed rock passes into unmetamorphosed rock.

Since the specific gravities and chemical analyses of the predominating silicates—garnet, epidote and ilvaite—roughly correspond, the errors in quantitative estimates of these minerals are not important for most of the oxides in the metamorphic zone. Alumina is the chief exception, for it is abundant only in epidote. The estimate for sphalerite is close because the quantity of zinc has been determined by stoping and development and assays of mill heads. Errors in the estimation of the magnetite result in corresponding errors for the iron oxides. An analysis of the ore was computed from available partial analyses of the ore which has been mined to date. When assaying, the insoluble silicates were not broken down; therefore, a large percentage of each analysis was reported as insoluble. This was divided by estimate into andradite, epidote, and salite. After the complete analyses of all the rocks and important minerals were made and the specific gravities determined, the quantities of the elements as oxides in the metamorphosed zone and those in the same block of ground before metamorphism were computed in metric tons. The volume of ground used for the measurements was 4,468,000 cubic meters contained in a block 383 meters long with an average cross section of 11,666 square meters. The block is bounded on the southeast by unaltered limestone

Material	S. G.	Vol. %	Metric Tons	SiO_2		Al_2O_3		Fe_2O_3		
				%	Tons	%	Tons	%	Tons	
Sill	2.73	3.26	397,644	58.89	234,173	16.55	65,810	0.55	2,187	
Shale	2.79	13.86	1,727,749	76.29	1,318,100	10.54	182,105	1.67	28,853	
Blue Marble	2.76	54.00	6,659,107	8.20	546,047	2.11	140,507			
White Marble	2.71	27.70	3,353,994	0.44	14,758					
Pre-Ore Dike	2.735	1.12	136,865	64.15	87,799	16.45	22,514	1.13	1,547	
	2.75	99.94	12,275,359	17.93	2,200,877	3.35	410,936	0.27	32,587	
Ore	3.20	4.02	574,765	20.20	116,103	0.79	4,541	22.62	130,012	
Low-Grade Ore	3.20	3.93	561,894	22.22	124,853	0.87	4,888	24.88	139,799	
Hornstone	2.79	8.04	1,002,243	76.29	764,611	10.54	105,636	1.67	16,737	
Epidote	3.35	15.40	2,305,041	36.48	840,879	21.37	492,587	12.46	287,208	
Garnet	3.84	57.70	9,899,658	36.27	3,590,606	0.51	50,488	29.97	2,966,928	
Salite	3.50	4.10	641,158	48.52	311,090	1.67	10,707	1.86	11,926	
Ilvaite	4.01	1.15	206,042	29.16	60,082	0.23	474	20.28	41,785	
Magnetite	5.17	4.83	1,115,707					69.00	769,838	
Chlorite	3.02	0.66	89,057	38.95	34,688	18.13	16,146	5.07	4,515	
Pyrrhotite, Pyrite	5.00	0.16	35,745							(Fe)
	3.68	99.99	16,431,310	35.56	5,842,912	4.17	685,467	26.59	4,368,748	
Metric Tons Gain or Loss			+4,155,951		+3,642,035		+274,531		+4,336,161	
Percentage Gain or Loss	+33.9	Assumed to be constant	+33.9		+165.5		+66.8		13,306.4	

* Adjusted

** Derived

Metric Tons	SiO_2		Al_2O_3		
	%	Tons	%	Tons	%
397,644	58.89	234,173	16.55	65,810	0.5
727,749	76.29	1,318,100	10.54	182,105	1.6
659,107	8.20	546,047	2.11	140,507	
353,994	0.44	14,758			
136,865	64.15	87,799	16.45	22,514	1.1
275,359	17.93	2,200,877	3.35	410,936	0.2
574,765	20.20	116,103	0.79	4,541	22.62
561,894	22.22	124,853	0.87	4,888	24.88
002,243	76.29	764,611	10.54	105,636	1.67
305,041	36.48	840,879	21.37	492,587	12.46
399,658	36.27	3,590,606	0.51	50,488	29.97
341,158	48.52	311,090	1.67	10,707	1.86
206,042	29.16	60,082	0.23	474	20.28
115,707					69.00
89,057	38.95	34,688	18.13	16,146	5.07
35,745					
131,310	35.56	5,842,912	4.17	685,467	26.59
55,951		+3,642,035		+274,531	
+33.9		+165.5		+66.8	

ved

TABLE 8.—*Tabulation of gains and losses of oxides in the metamorphic zone*

Original rocks

	Fe_2O_3		FeO		MgO		CaO		Na_2O		K_2O		H_2O+	
Tons	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons
35,810	0.55	2,187	4.56	18,132	2.72	10.816	7.60	30,221	3.32	13,202	2.00	7,953	1.49	5,925
32,105	1.67	28,853	4.71	81,377	1.19	20,560	0.10	1,728	0.17	2,937	1.45	25,052	2.63	45,440
0,507			0.91	60,593	0.10	6,659	*50.00	3,329,554						
			0.27	9,056	0.19	6,373	*55.30	1,854,759						
2,514	1.13	1,547	2.12	2,902	2.05	2,806	7.09	9,704	2.10	2,874	1.66	2,272	1.40	1,916
0,936	0.27	32,587	1.40	172,065	0.38	47,214	42.57	5,225,966	0.15	19,013	0.29	35,277	0.43	53,281

Metamorphic products

4,541	22.62	130,012	11.87	68,225	0.63	3,621	18.87	108,458			0.12	690	0.14	805
4,888	24.88	139,799	13.06	73,383	0.69	3,877	20.75	116,593			0.13	730	0.15	843
5,636	1.67	16,737	4.71	47,206	1.19	11,927	0.10	1,002	0.17	1,704	1.45	14,533	2.63	26,359
2,587	12.46	287,208	1.32	30,427	0.34	7,837	21.77	501,807	0.06	1,383	0.35	8,068	2.07	47,714
0,488	29.97	2,966,928	0.57	56,428	0.14	13,860	31.61	3,129,282						
0,707	1.86	11,926	18.56	118,999	4.11	26,352	21.71	139,195					0.75	4,809
474	20.28	41,785	31.06	63,997	0.67	1,380	13.29	27,383					2.19	4,512
	69.00	769,838	31.00	345,869										
6,146	5.07	4,515	21.50	19,147	4.06	3,616	2.09	1,861	0.05	45	0.53	472	7.76	6,911
			(Fe) 57.40	(Fe) 20,518										
5,467	26.59	4,368,748	5.14	844,199	0.44	72,470	24.50	4,025,581	0.02	3,132	0.15	24,493	0.56	91,953

Gains and losses

,531	+4,336,161	+672,134	+25,256	−1,200,385	−15,881	−10,784	+38,672
66.8	13,306.4	+390.6	+53.5	−23.0	−83.5	−30.6	+72.6

TABLE 8.—*Tabulation of gains and losses of oxides in the metamorphic zone of the Pewabic mine area*

Original rocks

	MgO		CaO		Na_2O		K_2O		H_2O+		H_2O-		CO_2		TiO
ns	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons	%
3,132	2.72	10.816	7.60	30,221	3.32	13,202	2.00	7,953	1.49	5,925	0.33	1,312	0.45	1,789	0.74
1,377	1.19	20,560	0.10	1,728	0.17	2,937	1.45	25,052	2.63	45,440	0.17	2,937			0.93
0,598	0.10	6,659	*50.00	3,329,554									** 39.41	2,624,354	
0,056	0.19	6,373	*55.30	1,854,759									** 43.66	1,464,354	
2,902	2.05	2,806	7.09	9,704	2.10	2,874	1.66	2,272	1.40	1,916	0.44	602	0.13	178	0.48
2,065	0.38	47,214	42.57	5,225,966	0.15	19,013	0.29	35,277	0.43	53,281	0.04	4,851	33.32	4,090,675	0.16

Metamorphic products

3,225	0.63	3,621	18.87	108,458			0.12	690	0.14	805	0.03	172	3.96	22,761	0.04
3,383	0.69	3,877	20.75	116,593			0.13	730	0.15	843	0.03	169	4.35	24,442	0.04
7,206	1.19	11,927	0.10	1,002	0.17	1,704	1.45	14,533	2.63	26,359	0.17	1,704			0.93
0,427	0.34	7,837	21.77	501,807	0.06	1,383	0.35	8,068	2.07	47,714	0.33	7,607	1.21	27,891	1.66
5,428	0.14	13,860	31.61	3,129,282							0.09	8,910			0.01
3,999	4.11	26,352	21.71	139,195					0.75	4,809	0.24	1,539			0.08
3,997	0.67	1,380	13.29	27,383					2.19	4,512	0.07	144			0.01
5,869															
0,147	4.06	3,616	2.09	1,861	0.05	45	0.53	472	7.76	6,911	0.25	223			0.72
0,518															
5,199	0.44	72,470	24.50	4,025,581	0.02	3,132	0.15	24,493	0.56	91,953	0.12	20,468	0.46	75,094	0.31

Gains and losses

,134	+25,256	−1,200,385	−15,881	−10,784	+38,672	+15,617	−4,015,581	+30
90.6	+53.5	−23.0	−83.5	−30.6	+72.6	+321.9	−98.2	+1

	TiO_2		P_2O_5		S		MnO		BaO		Zn		Total	
	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons	%	Tons
39	0.74	2,943	0.40	1,591	0.33	1,312	0.12	477	0.15	596			100.20	398,439
	0.93	16,068	0.04	691			0.11	1,901					100.00	1,727,749
54													100.73	6,707,719
54													99.86	3,349,300
78	0.48	657	0.24	328			0.20	274	0.25	342			99.89	36,715
75	0.16	19,668	0.02	2,610	0.01	1,312	0.02	2,652	0.01	938			100.36	12,319,922
61	0.04	230			7.72	44,372	1.58	9,081			11.06	63,569	99.63	572,640
2	0.04	225			** 3.66	20,565	1.74	9,777			5.50	30,904	98.07	551,048
	0.93	9,321	0.04	401			0.11	1,102					100.00	1,002,243
1	1.66	38,264	0.05	1,153			0.27	6,224					99.74	2,299,049
	0.01	990					0.98	97,017					100.15	9,914,509
	0.08	513					2.66	17,055					100.16	642,185
	0.01	21					3.13	6,449					100.09	206,227
													100.00	1,115,707
	0.72	641	0.34	303			0.48	427					99.93	88,995
					42.60	15,227							100.00	35,745
4	0.31	50,205	0.01	1,857	0.49	80,164	0.90	147,132			0.57	94,473	99.98	16,428,348

1	+30,537	−753	+78,852	+144,480	−938	+94,473	+4,108,426
2	+155.3	−28.8	+6,010.1	+5,448.0	−∞	+∞	+33.4

and sill rock, on the northwest by Hanover intrusive, on the northeast and southwest by extensions of the metamorphic zone, above by the surface, and below by unaltered limestone and hornstonized shale. The lower boundary closely corresponds to the main thrust fault and dips

TABLE 9.—*Chemical analyses of Hanover area dolomites and limestones*

	* DDH 112 107–177 feet (Middle Blue limestone)	* DDH 112 219–250 feet (Hanover limestone)	** Upper Blue limestone from surface west of Buckhorn Gulch	** Underground sample from middle of Hanover limestone	** Lower Blue limestone from 4th level	† El Paso limestone (Lower)	† El Paso limestone (Upper)	† Montoya limestone (Lower)	† Montoya limestone (Upper)
Insol.	11.52	0.49	16.8	††15.6	26.0				
SiO_2	8.2	0.44				20.62	8.57	34.03	2.77
Al_2O_3	2.11	none							
Oxides of Fe, Al, P, and Ti						7.59	3.45	3.96	1.76
Fe_2O_3									
FeO	0.91	0.27							
MgO	0.10	0.19	0.14	0.44	0.44	7.09	9.38	13.27	19.80
CaO	51.1	56.2	43.6	45.4	37.5	31.44	37.65	19.98	29.30
CO_2						32.51	39.60	28.67	46.01
Totals						99.25	98.65	99.91	99.64

* I. L. Wright, analyst—Diamond drill hole core from Pewabic mine.
** I. L. Wright, analyst—Hanover mine (Empire Zinc Company).
† J. J. Jones, analyst—Rocks below Percha shale.
†† Mainly chert.

flatly northwest (Pl. 2). The volume per cent of each rock, on the one hand, or each mineral or ore on the other, was multiplied by the individual specific gravity, the cubic meters of the metamorphosed block of ground, and the weight per cent of the oxides shown by chemical analysis of each rock or mineral. The net gains or losses were then calculated (Table 8). There were apparently large gains in silica (3,642,000 tons), ferrous and ferric iron (5,008,000 tons), sulphur, and zinc. There were moderate gains in alumina, magnesia, manganese, water, and titanium. There was an almost total loss of the carbon dioxide (4,015,000 tons) and a large loss of lime. Most of the soda and much of the potash

disappeared. The average increase in specific gravity was almost exactly one-third, which meant an average net gain of a metric ton of material to each cubic meter of original rock. The total gain of material per metric ton was slightly more than 2 metric tons, the total loss a little

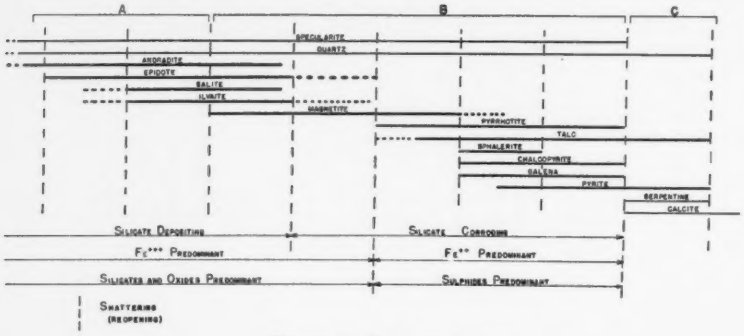


FIGURE 4.—*Paragenesis*

more than a ton. Alterations within the block of metamorphosed ground studied meant a local transfer of 2,230,500 metric tons of silica chiefly to form andradite. The net gain of silica in the zone is slightly greater.

The question of the origin and mode of transportation of the silica is fundamental. Since the commonest and quantitatively most important alterations in the district—epidotization, sericitization, and chloritization —imply a loss in silica and affect igneous rocks and shales which constitute probably half the rock fabric of the areas in and around the intrusive centers, the possibility that much if not all of the epigenetic silica could be derived from these alterations should be considered.

The net gain in silica would be reduced substantially if the endomorphic epidotization and chloritization of the adjacent part of the Hanover intrusive could be included in the balance sheet. But it is difficult to measure this alteration because of the lack of mine openings and drill holes. At the surface one area of epidotization extends for an average of 150 feet into the Hanover intrusive and is about 600 feet long. This would account for a loss of about 720,000 metric tons of silica if a possible extension under the intrusive is assumed to be equal to the surface length. This is equivalent to 20 per cent of the net gain of silica in the exomorphic zone. If one can believe that additional silica was derived from deeper-seated alterations, particularly epidotizations and chloritizations of shales and igneous rocks, including the main intrusive itself, one needs to account chiefly for the introduction of iron, the other metals and sulphur, and the extraction of lime and carbon dioxide.

It is of probable interest to compare the relative magnitudes of certain oxides in the Hanover intrusive with the same ones in the zone of exomorphism. The southern lobe of the Hanover intrusive may be assumed to be at least as large as an inverted cone 4000 feet in diameter and 750 feet deep at the center (Fig. 6) which calculates to 89,000,000 cubic meters or 240,000,000 metric tons.

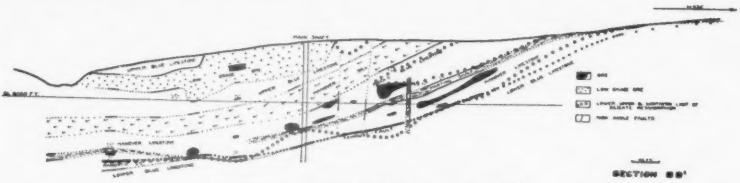


FIGURE 5.—*Longitudinal section BB'*
See Figure 3 for location.

The Pewabic mine part of the zone of exomorphism (Pl. 2b) is 1260 feet long and the balance, largely on the property of the Empire Zinc Company, is 3400 feet long. The cross section and composition of the zone do not vary greatly from place to place. The estimated net gain in amounts of certain oxides in the entire exomorphic zone, and, if determined, the per cent of similar oxides in the intrusive rock, and the relative per cent of exomorphic oxides gained compared to the mass of the intrusive rock are given in Table 10.

This comparison shows that the amounts of oxides and elements introduced into the exomorphic zone are, excluding silica and titanium, of a higher order of magnitude than the corresponding oxides and elements now in the intrusive. Zinc and sulphur were not determined in the intru-

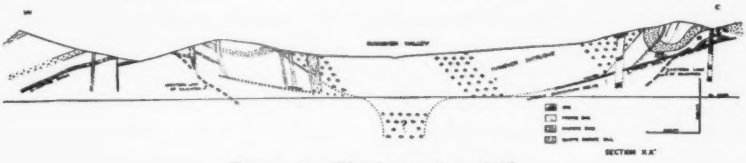


FIGURE 6.—*District section XX'*
An east-west section through southern part of Hanover intrusive.

sive. Clarke and Washington (1924, p. 23) state that there is an average of .00638 per cent ZnO (.00512% Zn) in a composite sample from 329 igneous rocks in the United States.

If we assume that the silica and iron gained by the exomorphic zone were derived from the adjacent mass of intrusive the original magma should have had more than 75 per cent SiO_2 and more than 11 per cent iron oxide. Certain basic rocks have as much silica but lower iron. The

inference seems strong that much if not most of the exomorphic oxides and probably the endomorphic iron oxide came from a source below the Hanover intrusive. If these oxides came from a deeper mass of magma the magnitudes of oxides in the intrusive compared with those of the oxides gained in the exomorphic zone suggest that this mass would have

TABLE 10.—*Estimated amounts of oxides and elements gained by the zone of exomorphism compared with amounts now in Hanover intrusive*

Oxides and elements	Metric tons gained in entire zone of exomorphism	Per cent of similar oxides in main intrusive	Per cent of oxides gained in exomorphic zone compared with mass of intrusive
Fe_2O_3 and FeO	19,000,000	3.82	7.90
SiO_2	13,500,000	65.36	5.60
Zn	600,000	n.d.	.25
S	400,000	n.d.	.18
MnO	550,000	.04	.23
TiO_2	50,000	.45	.02

to be considerably larger than the Hanover intrusive in order to have served as a source.

PARAGENESIS

The writer is sympathetic with the growing skepticism toward current methods of determining paragenesis and preparing paragenetic charts. The apparent unreliability of such work is often the result of insufficient samples of macroscopic observations and microscopic preparations. The writer's experience suggests that in many areas, if sufficient samples are studied, definite broad paragenetic relationships can be established. Chief reliance usually must be placed on the veining of one mineral by another. Inasmuch as shattering of many ore deposits occurred numerous times during the formation of the deposit—if, indeed, the occurrence of such shattering does not predicate many ore deposits because the first open spaces are filled by the first mineral deposition, thus choking the channel (Hulin, 1929, p. 23)—it is possible to date the minerals approximately by the successive veinings. Anomalous relationships such as reversals and omissions can be explained perhaps by minor pulsations in the composition of the mineral depositing fluids through changes at the source, and by shattering causing dislocation and intersection of channels (short-circuiting) carrying fluids of slightly differing composition. Apparently important variations of what may be called the standard paragenetic sequences occur but, judging from the literature and the writer's experience, these seem to be rare.

The writer has studied paragenesis in detail at Goldfield, Nevada, Santa Barbara and the Veta Colorado, Chihuahua, Mexico, and the Hanover area in New Mexico. In these deposits the universal broad sequence is an early deposition of essentially sulphide-free quartz and/or silicates (Butler, 1927, p. 235), followed after shatterings by metallic minerals (Wagner, 1925, p. 284). Inter-mineral shatterings are characteristic. These features also seem to be typical of the majority of carefully studied areas described in the literature.

Figure 4 shows the paragenetic relationships in the Pewabic area. The sequences were determined by numerous macroscopic observations of veinings, and 24 thin and polished sections were studied. The broader sequences such as the A, B, and C stages can be seen macroscopically. Andradite, epidote, and salite are cut by veinlets of sulphides and are often brecciated with the sulphides matrical to them. Microscopic study verifies the sequence of stages and places ilvaite as always pre-sulphides. Salite and epidote occasionally cut and replace garnet. The sulphides appear to replace the silicates, including the garnet. (*See* footnote 5.) In a few instances, magnetite is cut by veinlets of sulphides, and rarely by the early silicates, but much of it appears intergrown, and therefore contemporaneous, with the sulphides. It has never been observed to cut sulphides. In one instance it was cut and replaced by pyrite. Ilvaite is commonly cut and replaced by magnetite and sulphides but never cuts sulphides. Pyrrhotite is cut and replaced by other sulphides, sometimes is intergrown with them, and occasionally cuts them. Small amounts of specularite and quartz appear throughout the mineral sequence. Numerous shatterings (reopenings) during the mineralization epoch are conspicuous and significant.

The paragenetic relationships in the Pewabic area, although determined independently, do not differ much from those established earlier for the adjacent Empire Zinc Company property (Schmitt, 1935b, p. 197). The Pewabic mine study emphasizes even more strongly than the earlier study the early age of the high-temperature silicates and the late age of the sulphides. In both sequences the magnetite and ilvaite are intermediate in age, and the sequence is the same, *i.e.*, most of the ferric iron and silicon is early, and the ferrous iron, sulphur, and metals are late. The ilvaite and magnetite appropriately fall in the intermediate period when the deposition of ferric and ferrous iron overlaps (Butler, 1923, p. 403).

In the Pewabic area a separate study was made of the paragenesis in the three principal ore zones—Beta, Gamma, and Delta. In each there appears to be no essential qualitative difference in minerals or in their paragenesis. Macroscopic observations show, however, that there are

important quantitative differences in distribution which will be discussed under the next heading.

It should be noted here, for completeness, that in the Hanover intrusive a few feet from the contact with the silicate zone prominent veinlets of garnet and epidote cut aplite dikes. The aplite and pegmatite dikes are everywhere associated and even gradational in certain dikes. The pyrometasomatic stage is therefore later than the pegmatitic.

SPATIAL CONCENTRATION OF MINERALS

GENERAL STATEMENT

The differences in concentration of the various minerals appear to be caused by at least four controls, namely, differences in composition of the original wall rocks, in temperature of deposition, in time of deposition, and in structural forms. As previously mentioned the andradite and epidote of the first or "A" stage of deposition are largely in the broken and distorted ground above the basal thrust fault. Their relative distribution is closely governed by the original relative proportion and distribution of the aluminous (shales and dikes) and non-aluminous (limestone) wall rocks. Since the Pennsylvanian rocks overlying the Hanover sill are shaly and the sill is aluminous there is a zone with proportionately more epidote than andradite in the axial zone of the overturned syncline (Pl. 2, section AA′). Below the sill, andradite predominates except for the horizon of the Parting shale which is revealed as a sharply defined bed of epidote and which, incidentally, with the Hanover sill, reveals nearly all important details of the folding and faulting within the zone of metamorphism. The garnet extends slightly farther southeastward than the epidote which gives way to hornstone before the garnet yields to unreplaced marble. This may be because the hornstone was more difficult to penetrate and replace than the crystalline limestone or that the epidote was dependent on a higher temperature than the garnet.

The two remaining silicates, salite and ilvaite, of the late "A" stage are clearly segregated with the ore sulphides in a zone seldom more than 100 feet thick above the basal thrust fault. They are rare above this zone, except in the lens of low-grade ore which lies near the axial plane of the overturned syncline, and, like the ore sulphides, are localized by the northeast fissure zones. They overlap the garnet and epidote in time and, like the sulphides, appear to have been more dependent on open spaces for localization than the earlier silicates. The wall rocks appear to have been different also, for the fluids which deposited the garnet and epidote followed the same fissure zones as the later minerals but must have encountered easily replaceable limestone wall rocks, whereas the fluids which deposited the later silicates and magnetite, particularly

toward the end of their period must have encountered silicate walls which may have been relatively insoluble.

There is much more ilvaite, magnetite, and dark salite—about 10 to 1—on the intrusive side of the metamorphic zone than on the limestone side. Their largest concentration is in the southwestern part of the Delta zone northwest of the shaft. A large block of magnetite, and talc about 300 feet in diameter and 70 feet thick, with little admixed sphalerite and other sulphides, lies in the middle section of the mine north of the main shaft. The light coarse-grained salite is commonest on the limestone side of the metamorphic zone, particularly southeast of the main channel of the Beta fissure zone.

The sulphides and the latest (?) magnetite are confined to a zone overlying the basal thrust fault and, in more detail, to the three northeast-bearing fissure zones. There is a marked concentration of pyrrhotite in the Delta zone and, as previously noted, a greater concentration of chalcopyrite than is usual locally in the southwestern part of this zone on and just above the 370 level. The galena appears to be confined chiefly to the limestone side of the metamorphic zone as on the adjoining Empire Zinc Company property.

The late or "C" stage of the mineralization, when minor amounts of serpentine, pyrite, talc, and calcite were deposited, seems to be concentrated in the Delta zone, though confirmatory data are scarce.

Of the other minor minerals, wollastonite in small masses replaces chert on the limestone side, scapolite (marialite) is in the very edge of the Hanover intrusive, and chlorite (thuringite) is common in the pre-intrusive dike which outcrops above the Delta ore zone. Since the chlorite cuts epidote in the same dike and is near the intrusive it may belong to the "B" or "C" stage.

The close association of the gangue minerals, ilvaite, light and dark salite, and pyrrhotite with the sphalerite ore is one of the outstanding mineralogical features of the mine and is economically useful, for these minerals, when encountered in the walls, are seldom far from ore. The magnetite is more ubiquitous but, on the other hand, is seldom wholly free from disseminated sphalerite.

LOCALIZATION OF ORE BODIES

The distribution and relationships of the ore bodies of the Pewabic mine are shown on the composite plan (Pl. 2, a; see also various sections and plans). The shape of the ore is essentially that of connected horizontal pod-like to manto-like forms. Few ore bodies are more than 40 feet high, but several are 600 feet long. The two limestone beds previously described, particularly the Hanover limestone, enclose most of

the normal grade ore. In addition to this bedding control the ore is localized by other structural features (Pls. 2, 2, a, and Fig. 3) which include particularly the conspicuous pre-ore northeast vertical fissure zones Beta, Gamma, and Delta, the basal thrust fault, and especially the shear planes sympathetic to this fault and not far above it. These planes can be seen in some of the stopes and apparently cause the ore bodies, especially the pod-shaped ones, to cross-cut the favorable beds in planes approximately parallel to the thrust fault. Thus, in the horizontal zone much of the ore is positioned by the northeast fracture zones locally called channels, in the vertical zone by the favorable horizons, the basal thrust fault, and related flat shears. The pre-ore age of most of the vertical fractures is shown by their localization of veinlets of sphalerite and masses of silicates, of the thrust fault by the metamorphism of the associated gouge and breccia. There was some minor post-ore movement on the vertical fractures indicated by minor brecciation of the ore and deeply penetrating though restricted oxidation.

The typical ore body of the mine may be described as a pod-like to cigar-shaped mass about 40 feet in diameter, 600 feet long bearing northeast along strong northeast fractures, enclosed by the Middle Blue or Hanover limestone or both, and cross-cutting these horizons with a flat pitch generally not far above the basal thrust fault. In other words, it follows the intersection of the thrust fault and related flat shear with vertical fracture zones. On either side of the pod-like "back-bone", manto-like fins often spread out along cross fractures. These are likely to be less high but wider than the "back-bone" ore and may extend outward for several hundred feet.

The fact that relatively unimportant faults and associated fractures above the main thrust fault could localize large replacement ore bodies suggests that the mineralization intensity factor was great. In other parts of the Hanover district where silicates accompanying the ore are sparse, or lacking, large faults apparently are a requisite for ore localization even where the walls are the same favorable limestones found in the Pewabic mine. No ore is found on the smaller faults. This suggests that the mineralization was less intense at these places. The contrast is all the more significant when it is realized that in the Pewabic area, when the sphalerite stage had been reached by the mineral-depositing fluids, the walls were apparently no longer marble [5] but were silicates of varying solubility, presumably less soluble than marble under the

[5] This is indicated by the fact that unreplaced marble fragments have never been found well within the metamorphic silicate zone. Further, the early silicates are localized by the northeast fissures and extend along these fissures beyond the ore (Fig. 3). It seems, then, unlikely that limestone well within the outer edge of the zone where the ore bodies are now found could have escaped replacement by the early wave of silication along the fissures.

conditions that obtained. One asks why the "favorable" limestones were favorable to sulphide deposition. Several possible hypotheses appear. Throughout the Hanover area the silicates which replace the Hanover and Middle Blue limestones are coarser textured than those which replace the more aluminous beds above and below. It is well known that the coarser-grained aggregates of the silicates, particularly of garnet, can be drilled and blasted more readily than the fine-grained aggregates. Many specimens of garnet can be crumbled with the fingers. The inference is that the coarse silicates localized by the purer of the limestone horizons brecciated and crumbled more readily, and, since the age of the northeast fault zones is post-silicate in part, the silicates in the Hanover and Middle Blue limestones should be the better hosts for the ore sulphides. Another seemingly plausible hypothesis is that the proportion of the various silicates differs in the different horizons. Thus the "dirty" garnet in the shaly horizons is probably aluminous and more insoluble than the andradite in the Hanover limestone. The soluble ilvaite and magnetite are largely found in the Hanover and Middle Blue limestones. Deposition of sulphides would probably be promoted in the horizons containing the silicates most soluble under the prevailing conditions. The second process may work along with that first suggested related to texture.

There is a third hypothesis to consider. The thrust fault in the Pewabic area nearly always has the Hanover limestone as both walls, and always as one wall, which suggests that this limestone was more easily deformed and therefore selected by faulting. The localization of the ore in the Hanover and Middle Blue limestones could then be explained as due to the association of these beds with the thrust fault and not as due to their selective chemical influence on precipitation. This hypothesis, however, does not have much to commend it, for in all other parts of the district the Hanover and Middle Blue limestones, even though unassociated with thrust faults, are especially favorable to ore deposition. The writer favors the first two hypotheses and believes that the processes postulated were pertinent in influencing sulphide deposition in the Hanover and Middle Blue limestones.

The facts suggest that most of the sulphides were deposited in a silicate host. On the limestone side of the metamorphic zone, however, much ore has at least one wall of marble, and here probably the sulphides replaced the limestone directly. On the Empire Zinc Company property most of the known ore bodies have one silicate wall and one marble wall. Few ore bodies are known to occur well within the "contact metamorphic" zone. Even in the Pewabic area no important ore is known more than 150 feet from raw marble. Nevertheless, it seems as though we cannot

depend fully on the neutralizing power of the limestone to precipitate the sulphides within the silicate zone.

THEORY OF ORIGIN OF MINERALIZATION

It is pertinent now to review the facts and probable inferences bearing on the theory of origin of the mineralization, namely, those related to the probable path of the fluids which entered the block of ground here described, to the character of these fluids, to their change in character with time and flow, and to the deposition of minerals. The writer will not attempt to discuss alternate hypotheses in detail but assumes that possible discussion will present alternate views.

The scarcity of mineralized fractures or faults below the basal thrust fault, for only a few narrow garnet-filled veinlets are known below the fault east of the shaft, and the fact that the minerals formed at higher temperatures are most abundant on the intrusive side of the metamorphic zone suggest that a large proportion of the mineralizing fluids moved from the intrusive toward the limestone side in the block of broken, folded ground above the basal thrust and, on encountering the northeast fracture zones, were diverted upward.

The path of the probably gaseous, hotter and more penetrative fluids which deposited the silicates may have been more general and not so closely governed by structural features, particularly the thrust fault, as the path of the probably liquid fluids which deposited the sulphides. On the other hand, unaltered blocks of limestone begin to appear upward as the distance above the thrust fault increases (Pl. 2, section AA′), suggesting that the thrust fault zone acted as a principal channel which guided the silicate-bearing fluids. The deposition of both the silicates and sulphides was definitely controlled in part by the northeast fractures (Fig. 3), but presumably this deposition, particularly of the silicates, erased most of the evidence of control by structure. Thus, the evidence suggests that the fluids which deposited both the silicate and sulphide stages of the pyrometasomatism moved from some intrusiveward channel [6] flatly outward and upward in and/or not far above the basal thrust fault. Details of the silicate and sulphide localization above this fault were controlled by smaller structural features.

The facts which suggest the character of the mineral-depositing fluids and their change with time and position are, of course, indirect and inconclusive. The fluids which deposited the early "A" stage (Fig. 4) silicates, chiefly andradite and epidote, apparently had the power, when pure limestone was the host and this was the chief rock metamorphosed, to break

[6] Possibly coinciding with a deep-cutting keel-like dike or volcanic neck marking the downward prolongation of the intrusive because all known contacts of the Hanover lobe dip inward.

down the $CaCO_3$, add and combine silica and ferric iron to sufficient lime to form pure andradite (Pl. 1, c), and to expel all the CO_2 and the remaining lime (Butler, 1923, p. 401). Incidentally, calcite and other carbonate material are found in veins in the outer mineralized zone of the district. Some lime was used to form epidote, but Table 8 indicates that some was exported. The early fluids in aluminous rocks such as shale, shaly limestone, and igneous rocks (Pl. 1, a, b), also had the power to oxidize the ferrous iron and add more ferric iron, lime, and alumina to form epidote. They subtracted silica and the alkalies.

In the late "A" stage salite, ilvaite, and magnetite were being deposited in addition to andradite and epidote. Apparently the early silicates were unstable and were replaced by the later silicates and magnetite, for macroscopic and microscopic veinlets of ilvaite and magnetite cut the early silicates, and in thin section euhedral crystals of magnetite apparently replace silicates. A magnetite crystal, for example, overlaps (replaces) the boundary area of two silicate crystals, a phenomenon which cannot be explained as interstitial magnetite in brecciated silicates. It is difficult to separate quantitatively and in detail the silicate and sulphide stages, but since by far the largest amount of material was supplied in or during the silicate stage, and the sulphide stage largely accounts for the metals other than iron and the sulphur, the balance for iron, lime, silica, alumina, and carbon dioxide as given on Table 8 is a close approximation for the early stage. Apparently, then, in the early stage one needs to account largely for the addition of iron, a large portion of the silica, and the subtraction of some of the lime and most of the carbon dioxide.

The fluids which deposited the sulphides apparently could replace [7] the early silicates and oxides including salite, ilvaite, and magnetite, and, although they could replace limestone directly, their character may have been changed by the first alterations before they reached this rock. In replacing early silicates they must have taken up and exported silica, iron, lime and probably alumina from the epidote, and magnesia from the salite. They deposited sulphur, zinc and other metals, and ferrous iron. They seem to have been strongly reducing in character because ferrous iron appears to have been stable, ferric iron unstable,[8] in the solid phase. A small amount of quartz but no silicates was deposited.

The fluids which effected the metamorphism, then, contrasted greatly

[7] The case for the replacement of early silicates by sulphides and other minerals seems as secure as the case for replacement of limestone by minerals. The sulphides are definitely later than the silicates yet occur as massive ore bodies well within the silicate zone. Presumably they have replaced the silicates. In thin section sphalerite apparently replaces massive crystalline salite and andradite. Some sulphide advanced as rounded replacement fronts across crystals of silicates.

[8] Many writers have noted the bleaching effect of sulphide-depositing fluids upon sandstones and volcanic rocks colored by ferric oxide and other ferric minerals.

from first to last, at least when they were depositing minerals in the metamorphic zone. The writer believes that a hot oxidizing, acid gas can best account for the deposition of the early minerals. Acid liquids should probably be eliminated because epidote, magnetite, and ilvaite are unstable in them at least under laboratory conditions. An alkaline liquid or gas should probably be eliminated because this doubtfully could react with or decompose the limestone unless the temperature were over 800 degrees, and the CO_2 back pressure would increase this. But the temperature does not appear to have exceeded 575 degrees because the quartz, which in places with specularite fills the interstices between andradite crystals, appears to have been formed as alpha quartz. Microscopically this quartz is clear and unbroken, and the crystals appear to be trapezohedral-tetartohedral in form. A conventional objection to the postulate that acid gases or liquids were carriers for metallic and other cations (Geiger, 1925) in the Hanover area would be the lack of large quantities of halogen-bearing minerals. There is some scapolite in the intrusive, a little apatite in the contact zone on the west side of the intrusive, and vesuvianite and fluorite in an area about 3000 feet west of the same contact. The amount of these minerals, however, falls short of what Geiger would place as a minimum for the large quantities of iron deposited. The objection may be difficult to surmount because scapolite, apatite, and fluorite (Fenner, 1933, p. 99-100), and possibly vesuvianite (Clarke, 1924, p. 276), are stable in acid gas environments. (However, *see* Zies, 1929, p. 70.) These four minerals are unstable in an acid liquid environment.

The fluid which deposited the sulphides was probably a weakly acid reducing liquid. It was probably a liquid because the magnetite and ilvaite were replaced (dissolved). Probably neither an acid nor alkaline gas could do this. It was presumably weakly acid because sulphides are deposited from such a medium in the laboratory and because raw limestone was attacked. An alkaline liquid would not have decomposed limestone unless the temperature were above 800 degrees for the reasons noted for alkaline gas. Neither would it have attacked and replaced the magnetite and ilvaite, for both are insoluble in alkaline liquids under laboratory conditions. The temperature must have been under 575 degrees since quartz was deposited with the sulphides, rather than garnet or other silicates, even though iron and lime were in solution. It has been shown that the early silicates were deposited below 575 degrees. After passing out of the pyrometasomatic zone along channels, the liquid possibly remained slightly acid for limestone was still actively attacked and replaced, and calcite was unstable along the principal channels until the outermost metallogenetic zones of the district were reached. There,

abundant calcite occurs in the veins. Along minor fissures and veins nearer the pyrometasomatic zone, calcite containing minor amounts of iron and manganese and other carbonate was deposited with no accompanying minerals.

There was much less silica deposited toward the last, and relatively more iron. This suggests that the emanations carrying metals, chiefly iron, lacked silica near their source and robbed some from the silicates in the walls along their channels of escape by replacing them mainly with epidote and chlorite which minerals at the same time tied up much of the iron. When the limestones were reached the silica with the remaining iron was combined with the lime to form garnet. Within the metamorphic zone additional silica was robbed from the shales and igneous rocks and transferred to the limestone beds. The deposition of the early silicates from the gas in the relatively restricted pyrometasomatic zone was no doubt partly or largely due to the neutralizing effect of the limestone. Temperature drop may have aided, although it is difficult to believe this could have been effective over so narrow a zone, particularly along open channels, and, further, the walls must have been heated up for several thousand feet from the silicate zone because hornstonization of the shales and recrystallization of the limestone extend at least for this distance.

When the channels along which the gases had "leached" silica were jacketed by epidote and chlorite the gases may have remained silica poor and relatively iron rich. When they reached the metamorphic zone, then, they may have been able to attack the earlier-deposited andradite and epidote and to deposit magnetite, salite, and some of the iron-rich ilvaite. Judging from the work on fumaroles by the geophysicists, it seems probable that lead and zinc cations were present in the early gases as well as in the later fluids but could not be deposited at the high temperatures prevailing at first. The marked increase in ferrous iron deposition with time suggests an increase in the reducing character of the fluids, a change which continued until, toward the end of the sulphide stage, no ferric iron, only ferrous iron, sulphur and metals were deposited. The data tend to support the idea of a chemical system continually changing in acidity and oxidizing power, but without marked steps. In summary the most applicable postulate is one of early acid, oxidizing gases which gradually became more reducing, less acid, and cooler until, at the end of the stage of magnetite deposition, which was before the end of the sulphide stage, slightly acid liquids which could attack magnetite and deposit sulphides prevailed. Finally the mineralizing liquid became alkaline when carbonates were stable in the solid phase.

The fact that much of the silica, now tied up as silicates which replace the limestone, can be explained as derived from the alteration of shales and igneous rocks (Knopf, 1929, p. 33; Paige, 1922, p. 28; Spencer, 1917, p. 64) seems significant. The abundance of silica in most ore deposits and the occurrence of metals in the pegmatites early led to the hypothesis of descent from magmas through pegmatites to veins. The scarcity of sulphides in pegmatites and/or the lack of clear linkage between the pegmatites and sulphide veins have been puzzling although a few linkages have perhaps been proved between pegmatites and massive quartz-gold veins. If pegmatites are accepted as one of the essential links, it is difficult to explain the shallow extension with depth of most ore deposits. A general appreciation of the magnitude of the silica derived from wall-rock alterations may swing the emphasis from pegmatites and place it where it seems more natural to some, *i.e.*, on the so-called gaseous volcanic emanations whose possibilities as carriers of the metals and silica from magmas are now well known. Such emanations can replace earlier minerals, drop certain elements, and pick up others. (*See* Allen, 1922a, 1922b; Allen and Zies, 1923; Allen and Day, 1924, 1927, 1928; Day and Shepherd, 1913; Shepherd, 1921, 1925; Shepherd and Merwin, 1927; Zies, 1929; Fenner, 1933.) Thus it appears that the origin of the Hanover mineralization can be traced back to volcanic exhalations if we can believe that the silica need not be accounted for as a magmatic differentiate and if we can conceive of it rather as having been derived from alterations along the path of the escaping gases, particularly by the metasomatism of shales and igneous rocks.

The objection of the geophysicists that liquids could not rise from a magma against the hydrostatic head (Day and Allen, 1924, p. 184-186) can probably be fitted into the postulated genetic pattern. During the early hot, explosive stage of volcanism, water could not exist as a liquid in the zone between the source of the gases and the surface, but, as the intrusive and volcanic activity died down, meteoric water could penetrate farther and farther downward. This would correspond to the hot-spring stage and would be equivalent to stages "B" and "C" of the Pewabic mine paragenesis. The meteoric water could probably absorb much of the metal-laden gas emanation and under certain conditions would drop some or most of this load.

There are no known data which give a good value for the depth at which the mineralization took place at Hanover. The age of the mineralization is post-Colorado and pre-Middle (?) Tertiary. The sulphide silicate gangue replacement is found stratigraphically at least as high as the base of the Beartooth quartzite. Quartz-sulphide veins cut the more than 2000 feet of Colorado sandstone and shale and intercalated

sills which lie above the quartzite. Spencer and Paige (1935) report still younger pre-mineralization (?) volcanic rocks of Cretaceous age in the northwest part of the Santa Rita quadrangle, but, since the thickness of these is unknown, the distance from the Hanover limestone to the surface in mineralization time can only be said to be greater than 3500 feet, or if we may assume that few volcanic piles in the southwest in Cretaceous time were likely to have been thicker than 5000 feet we might estimate the depth of origin at not more than 8500 feet.

Most economic geologists active in the field are impressed with the tendency of ore deposits to bottom at shallow depths (Hoover, 1909, p. 30-33). The hypothesis suggested above would demand such a condition because of the known moderate depth of penetration of meteoric waters. The objection that meteoric water does not appear to penetrate deeply enough in mining districts may be met with the argument that when an area is being mineralized fresh breaks and folds would permit water to penetrate deeply (Lindgren, 1935, p. 465). In the later less active stages openings would tend to remain sealed by deposition.

The evidence suggests the existence of a marked hypogene physical and chemical unconformity at shallow depth during the deposition of certain ore deposits. A rising hot, acid metal-bearing gaseous emanation at some time during the approach toward the quiescent stage of the associated volcanism must encounter descending meteoric water and be partly and finally completely absorbed. Progressive super-saturation of the meteoric water, neutralization of the acid, reduction of temperature, upward enlargement of channel, and exposure to much greater rock surfaces should result in relatively abrupt deposition of much of the load carried in solution. Many hypogene ore deposits show a marked concentration of the ore minerals in the horizontal zone, *i.e.*, have a limited extension vertically which suggests dumping.

SUMMARY OF CONCLUSIONS

1) The facts indicate that the southern lobe of the Hanover intrusive is floored and that the magma made room for itself by centrifugally thrusting aside its walls. The mechanism of the movement was essentially that of a basal thrust fault, whose upper plate was folded. The lower plate remained essentially undisturbed. In some places the folds are overturned and impressive rock flowage assisted in the adjustments.

2) The silicate and sulphide pyrometasomatism which includes some endomorphism is confined chiefly to a zone several hundred feet wide adjacent to the intrusive. It appears to be controlled by the proximity and structure of the intrusive contact and the structural forms associated with the thrust fault. The ore bodies are localized in detail by the coin-

cidence of vertical northeast fractures, the thrust fault, and favorable limestone beds.

3) A detailed balance of the gains and losses of the oxides in the metamorphic zone made possible by the numerous measurable exposures given by extensive mining, exploration, and diamond drilling shows that the principal gain was in the iron oxides, silica, zinc, manganese oxide, sulphur, and alumina listed in order of magnitude. The principal losses were in carbon dioxide and lime. The net gain in material was nearly one metric ton for each cubic meter of rock replaced. The total amount of iron oxide introduced into the metamorphic zone bordering the Hanover lobe of the intrusive was about 19,000,000 metric tons, of silica 13,-500,000, and of zinc 600,000 metric tons. The evidence favors constant volume during replacement.

4) The paragenetic relationships can be established macroscopically and microscopically. The silicates are early and have been replaced by magnetite and the sulphides along post-silicate faults and fractures. The early stage was marked by predominant ferric iron, the late by predominant ferrous iron. The paragenesis and replacement phenomena suggest marked physical and chemical changes with time in the fluids which deposited the minerals. The most likely agents for the transport and deposition of the early silicates and magnetite seem to have been acid gaseous volcanic emanations, for the late sulphide stage slightly acid liquids. The late and/or distant carbonates were probably deposited from alkaline solutions. The changes are considered to have occurred progressively with the waning of volcanism and invasion of the metalliferous zone by meteoric waters.

5) The hypothesis, not new, is here supported for the origin of some ore deposits from volcanic emanations with meteoric water taking an important rôle. This hypothesis seems reënforced in proportion as one believes that the silica associated with ore deposits is derived from alterations such as sericitization, chloritization, epidotization, and carbonatization. One is thus released from dependence on differentiation to account for the silica, and the magma need be called upon only to provide the metals, halogens, sulphur and water which are the principal elements of the volcanic exhalations.

In the Hanover area, at least, the net gain in silica seems reasonably accounted for as derived from the epidotization and chloritization of adjacent and deeper shales and igneous rocks including the intrusive itself. The net gain in iron oxide is larger than the net gain in silica. It is believed that quantitative studies in other areas will reveal similar relative magnitudes and subordinate the rôle of magmatic silica.

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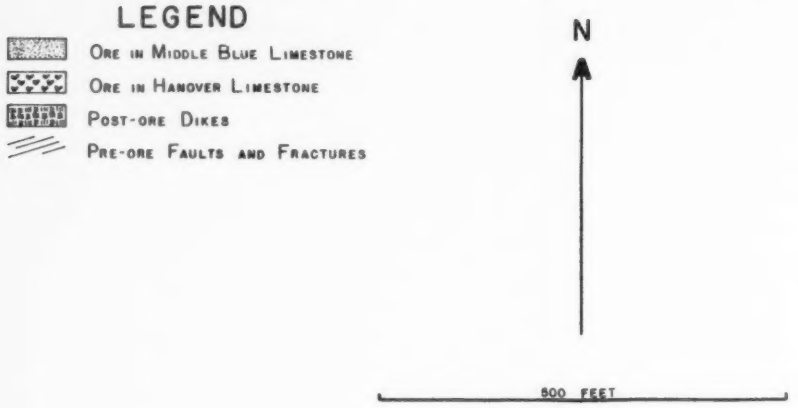
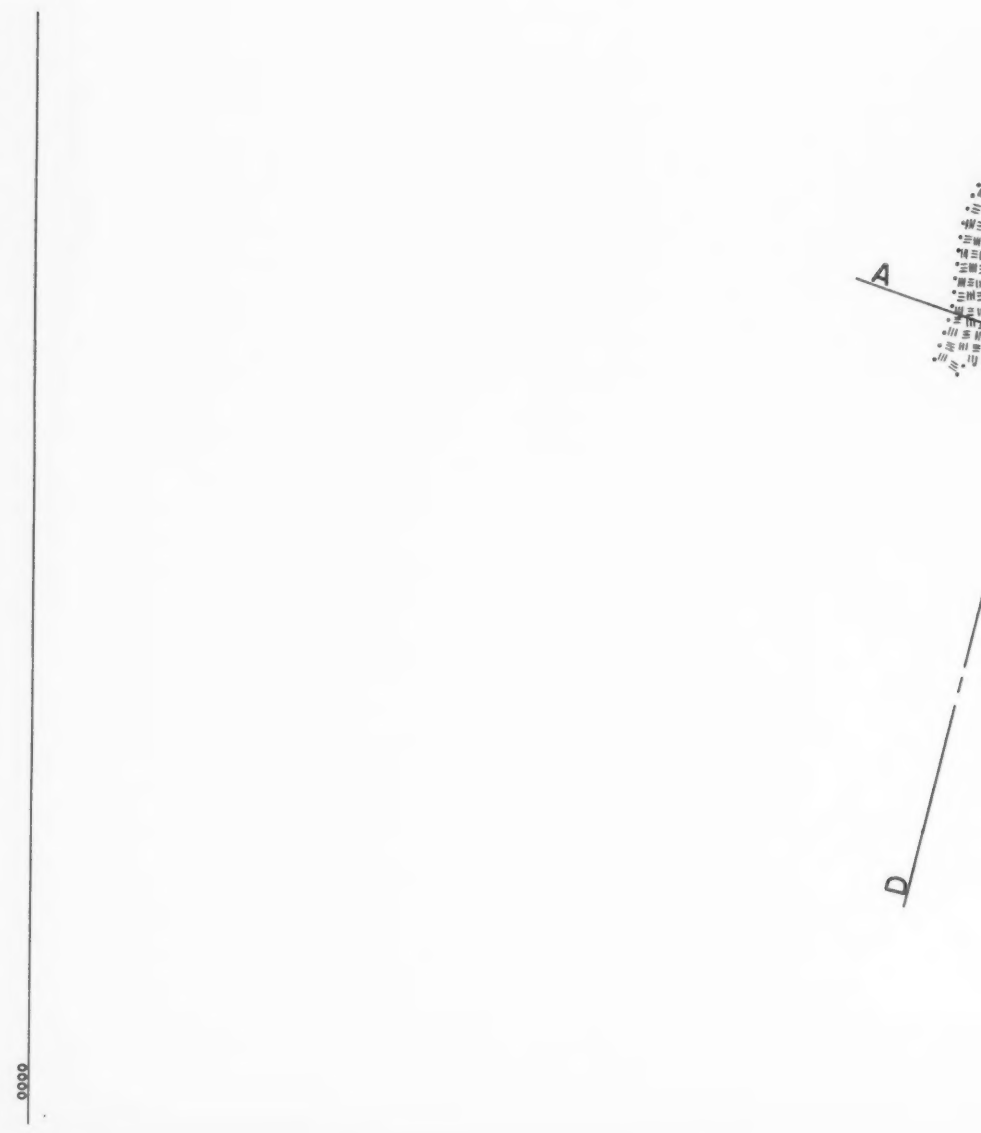
Read before the Geological Society of America, December 30, 1938.
Project Grant 143-35.





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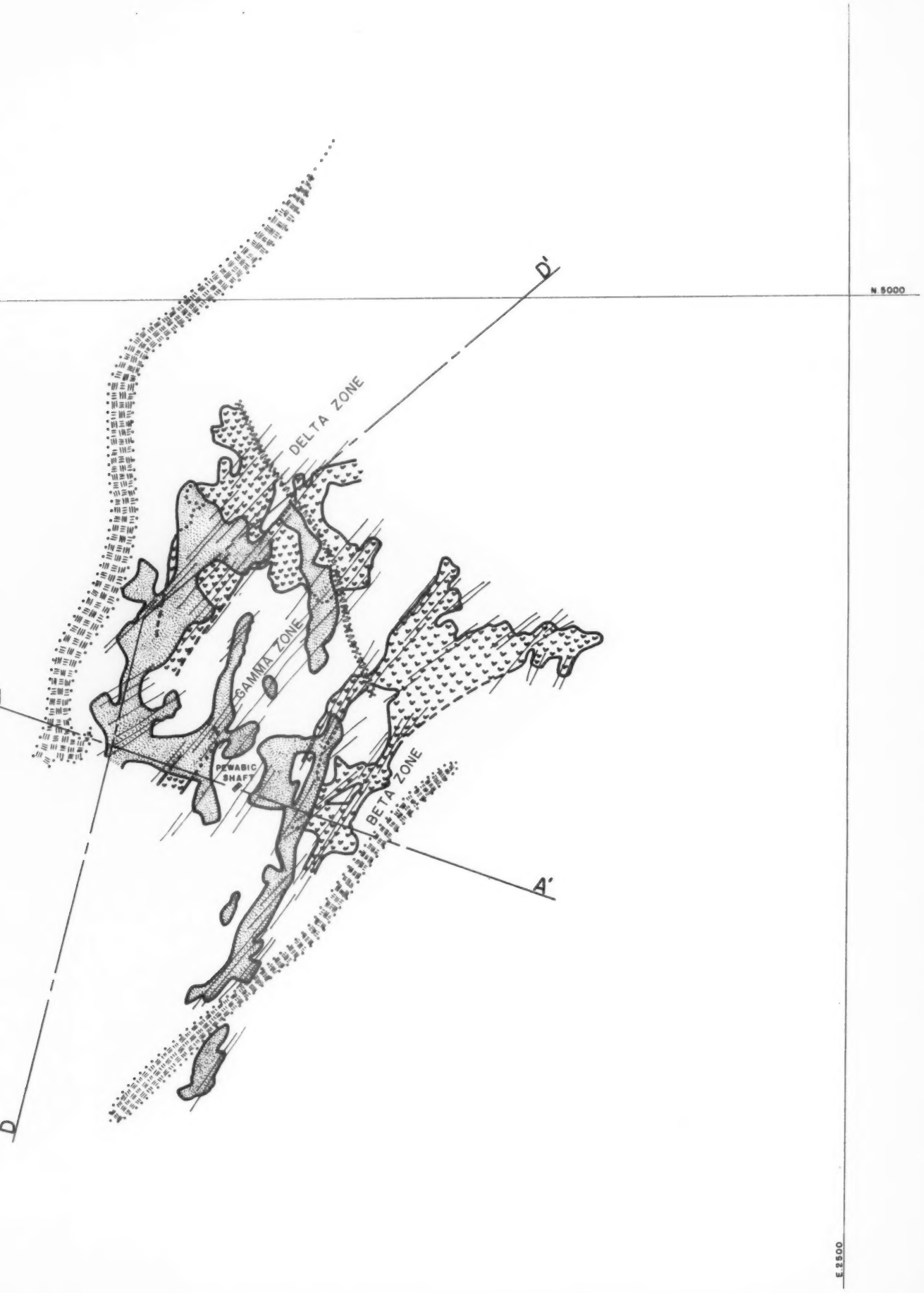
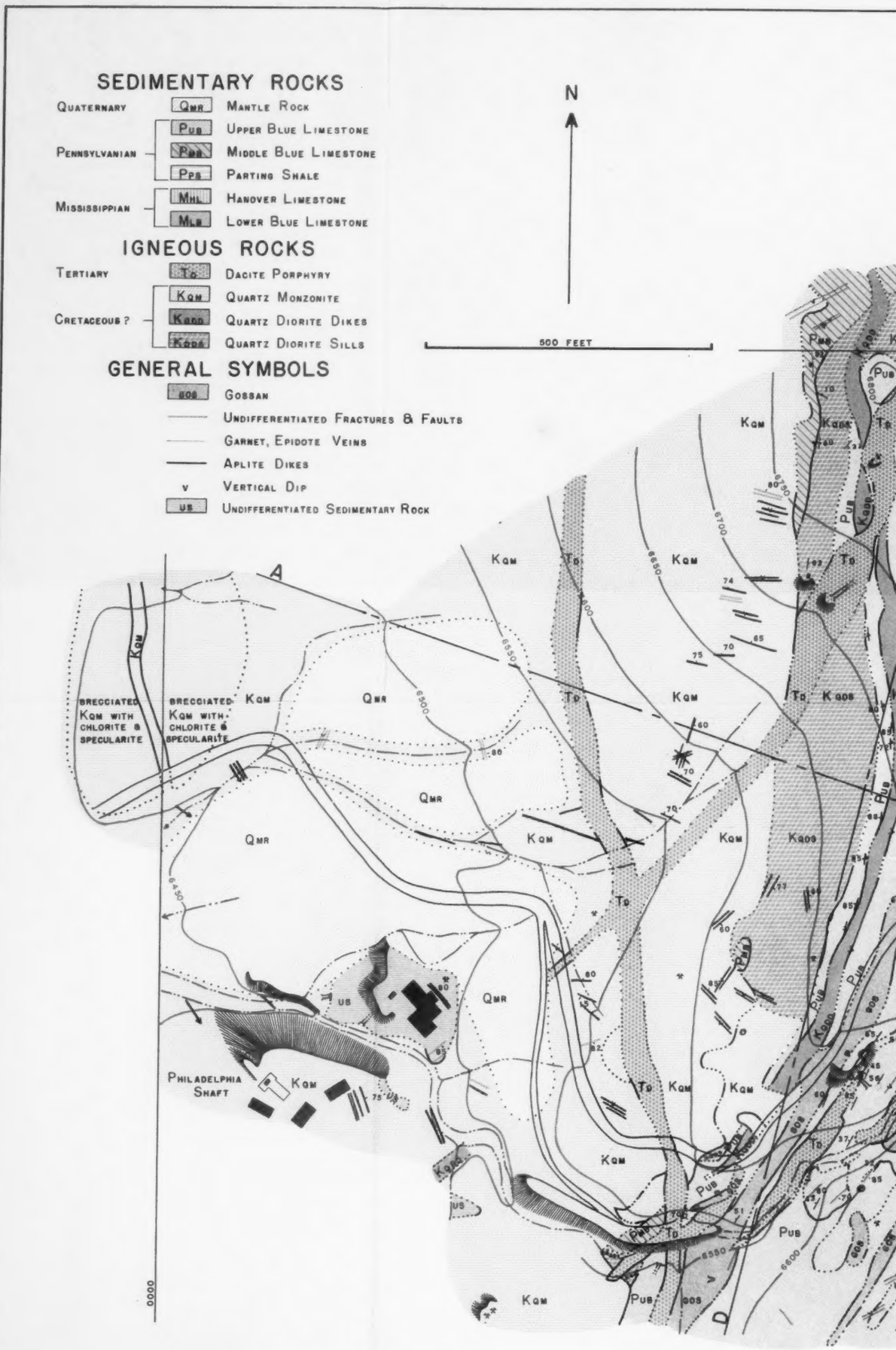


PLAN OF ORE BODIES

ENDOMORPHISM AND

AND EXOMORPHISM


SEDIMENTARY ROCKS
Quaternary
Qmr Mantle Rock
Pennsylvanian
Pub Upper Blue Limestone
Pmb Middle Blue Limestone
Pps Parting Shale
Mississippian
Mhl Hanover Limestone
Mlb Lower Blue Limestone
IGNEOUS ROCKS
Tertiary
Td Dacite Porphyry
Cretaceous ?
Kqm Quartz Monzonite
Kqdd Quartz Diorite Dikes
Kqds Quartz Diorite Sills
GENERAL SYMBOLS
gos Gossan
Undifferentiated Fractures & Faults
Garnet, Epidote Veins
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Brecciated Kqm with chlorite & specularite
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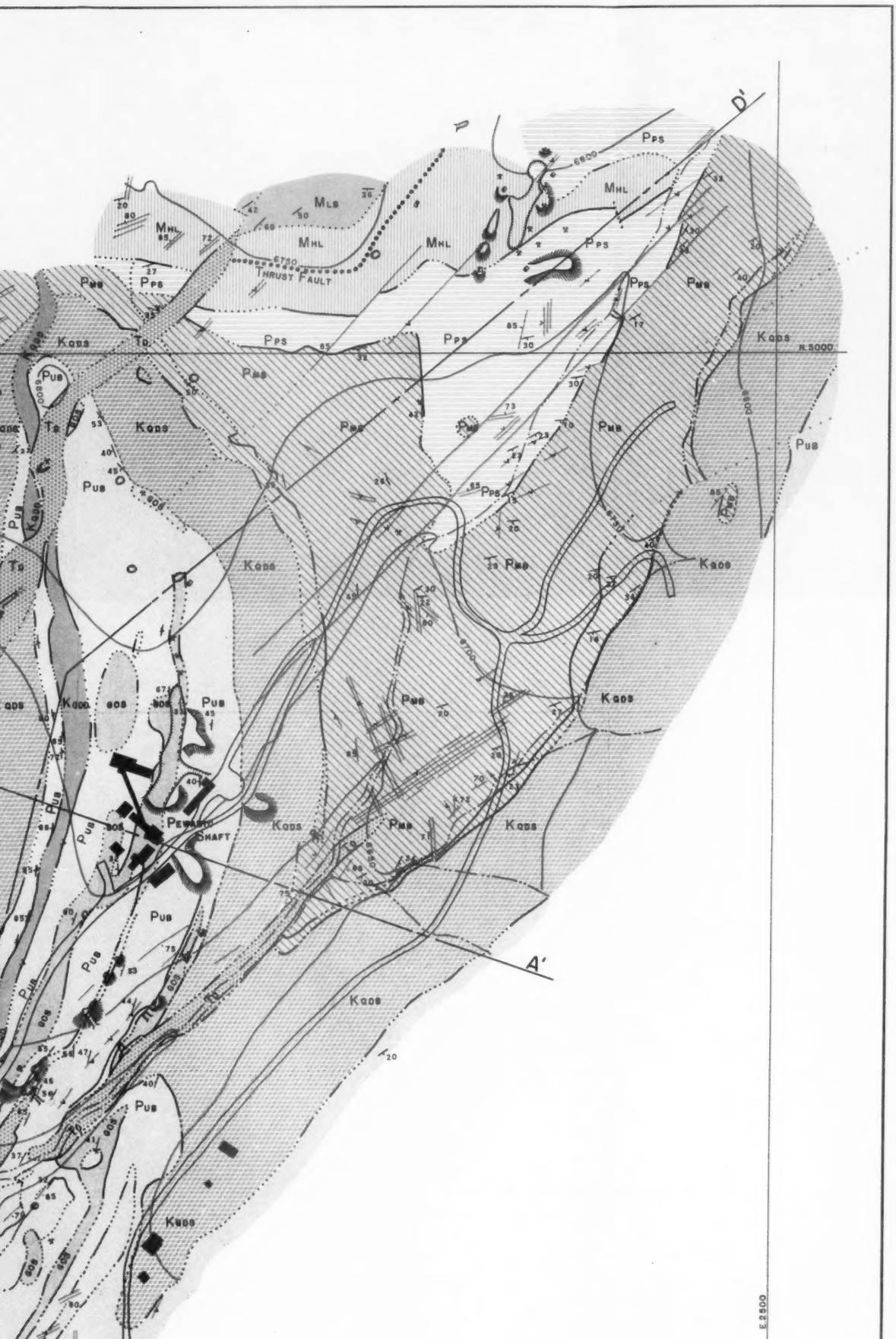
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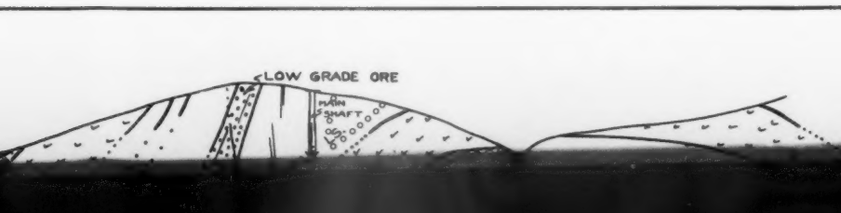
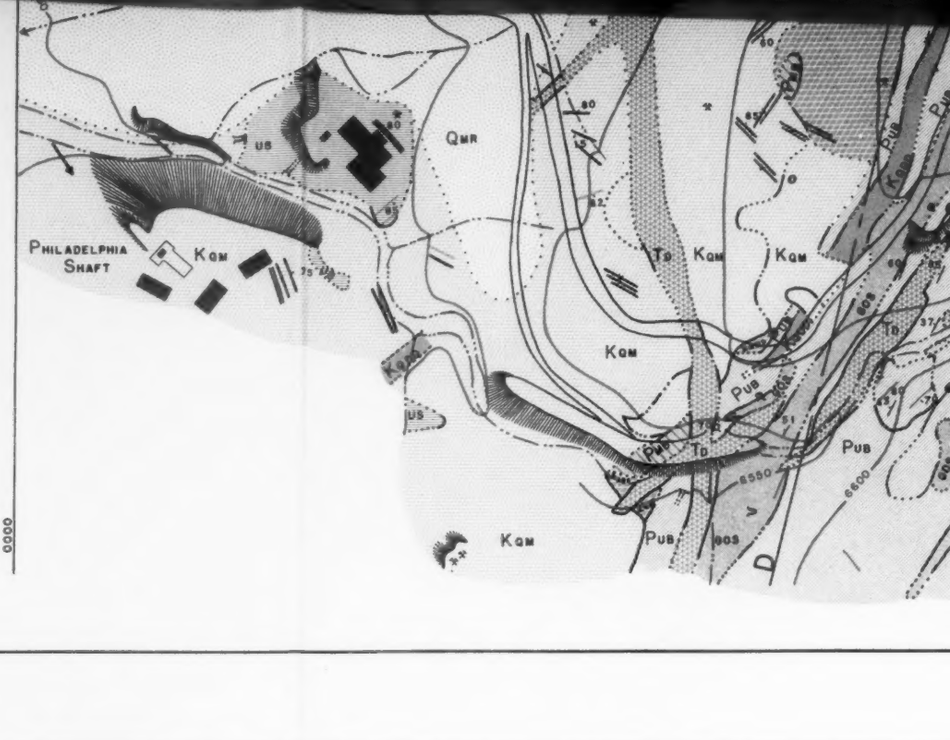
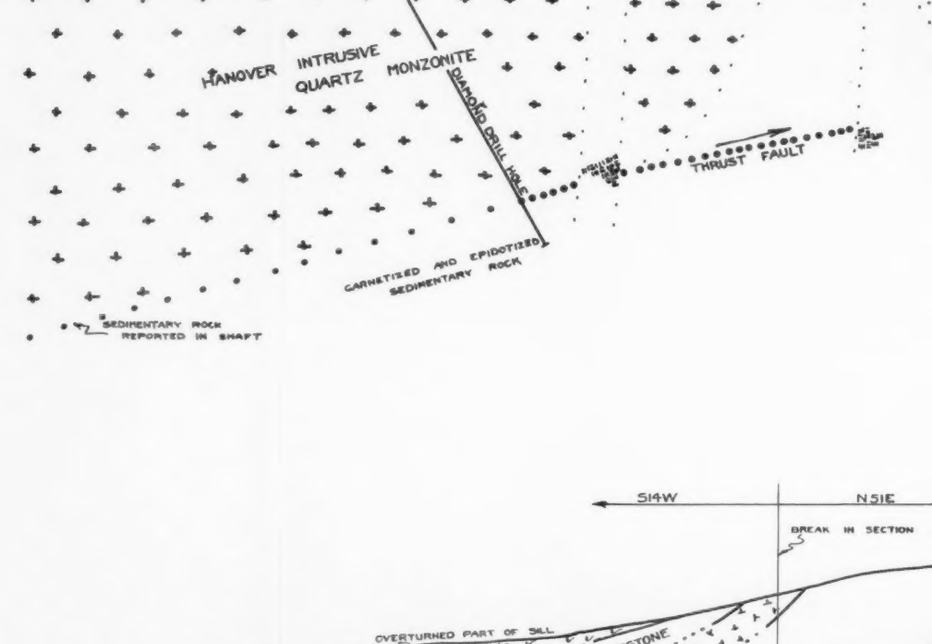
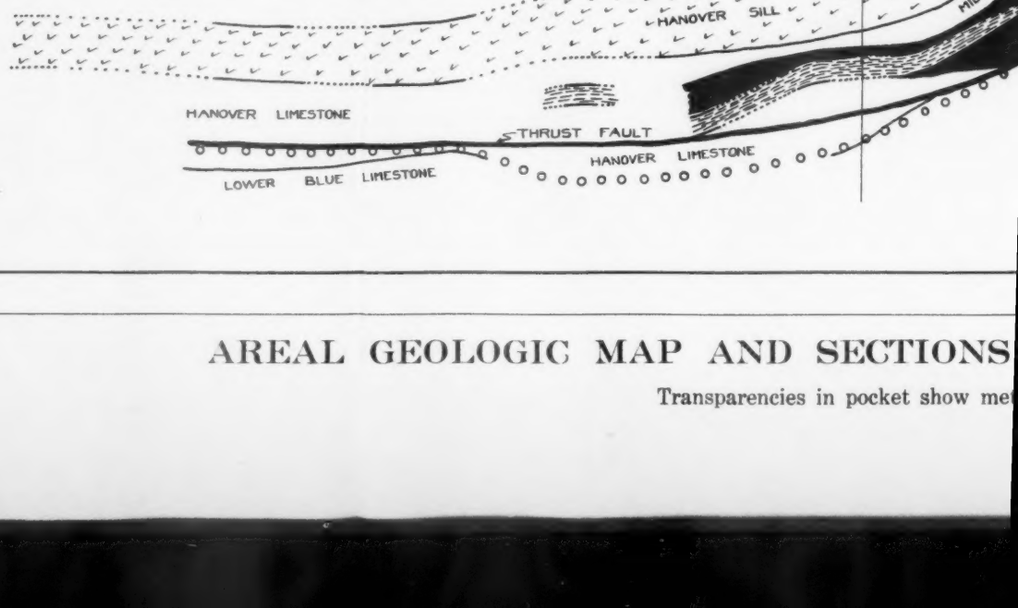

LOW GRADE ORE
PEWABIC SHAFT







AREAL GEOLOGIC MAP AND SECTIONS

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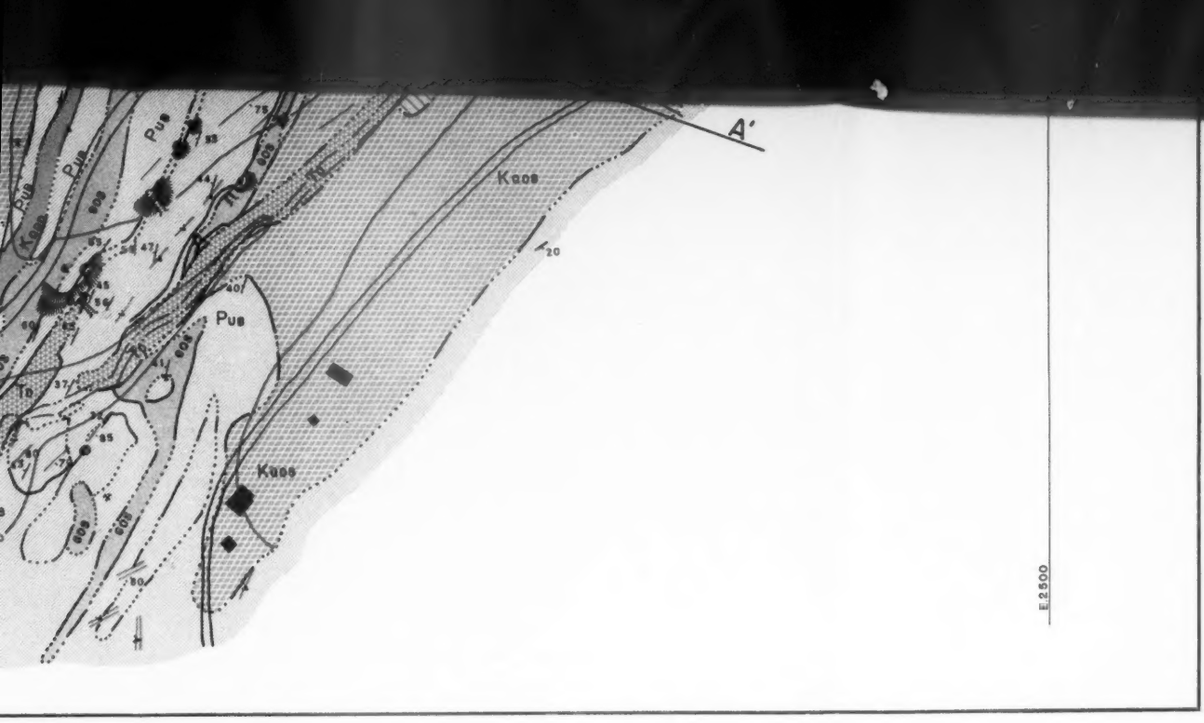

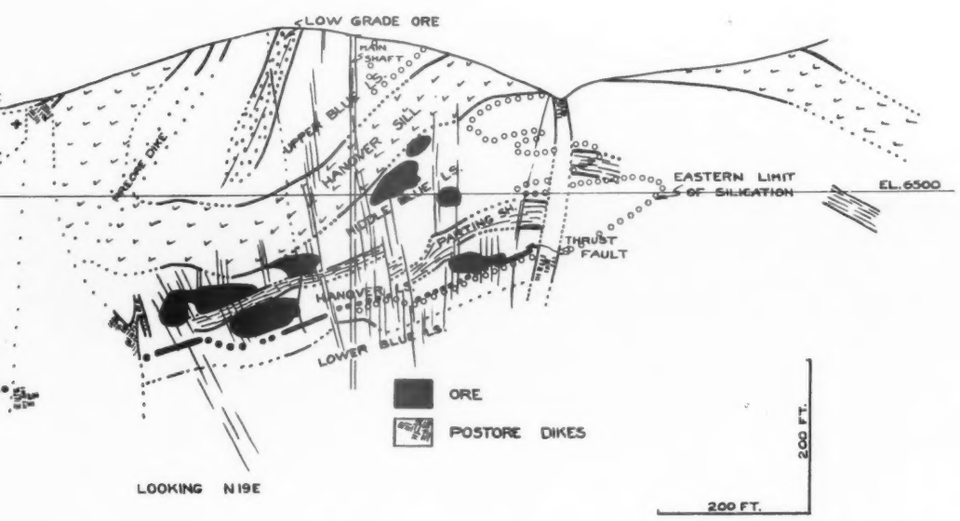

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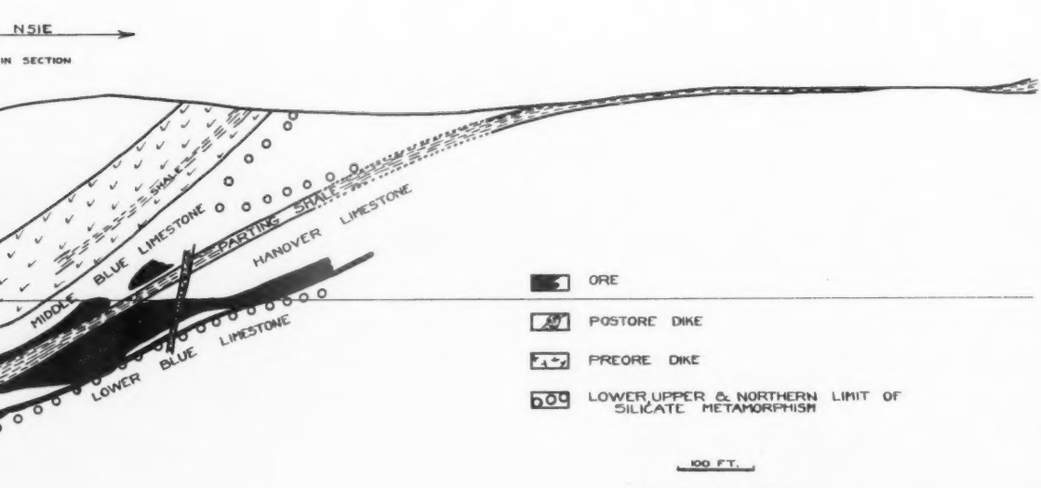

SECTION DD'

IONS OF PEWABIC MINE AND VICINITY

show metamorphism and ore bodies